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by

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SYMBOLS

A	Coefficient used in Eqn (14); defined in Eqn (15).
c	Used to denote a constant (e.g. in Eqns (15) and (46)).
c_g	Specific heat capacity of soil ($J\ kg^{-1}K^{-1}$).
c_p	Specific heat of air at constant pressure ($J\ kg^{-1}K^{-1}$).
C	Volumetric heat capacity of soil ($J\ m^{-3}K^{-1}$) (Used in Eqn (3) and Figure 12 only as denoting cloud amount).
C_{eff}	Effective thermal capacity of surface layer of soil ($J\ m^{-2}K^{-1}$).
C_ξ	Bulk transfer coefficient appropriate to the vertical surface turbulent flux of some conservative quantity whose mean value is ξ.
C_D, C_H, C_E	The appropriate bulk transfer coefficients for the surface turbulent fluxes of momentum, sensible heat and water vapour, respectively.
d_w	Representative depth of water in surface soil layer (i.e. the soil moisture content) for hydrological calculations (m).
d_s	Snow depth (m).
d_{sw}	Water equivalent depth of snow ($d_{sw} = 0.1\ d_s$) (m).
D	Effective depth of soil assumed to determine the effective thermal capacity of surface layer of soil (m).
E	Surface flux of water vapour; surface evaporation rate; surface evapotranspiration rate ($kg\ m^{-2}s^{-1}$).
E_p	Surface potential evaporation rate ($kg\ m^{-2}s^{-1}$).
E_s	Surface sublimation rate ($kg\ m^{-2}\ s^{-1}$).
F	Surface infiltration rate ($kg\ m^{-2}\ s^{-1}$).
F_ξ	Mean vertical surface turbulent flux of some conservative quantity whose mean value is ξ. Units are those of $(w\xi)$.
g	Acceleration due to gravity ($m\ s^{-2}$).
G, G_o, G_i	Vertical heat flux in the soil; G_o is the value at the surface; G_i ($i = 1(1)4$) denotes values at specified depths in the soil (Wm^{-2}).
H	Surface sensible heat flux (Wm^{-2}).

k	von Karman constant (≈ 0.4).
K, K_s	Hydraulic conductivity of the soil; K_s is the saturated conductivity ($m\ s^{-1}$).
L	Monin-Obukhov length; $$L = - \frac{\rho c_p T \quad u_*^3}{kg[H+0.61c_pTE]} \text{ (m)}.$$
L_i, L_e, L_f, L_s	L_i is one of: the latent heat of evaporation, L_e; the latent heat of fusion, L_f; the latent heat of sublimation, $L_s = L_e + L_f$ ($J\ kg^{-1}$).
m_w	Mass of liquid water per unit lateral area in a soil layer of specified depth ($kg\ m^{-2}$).
m_s	Mass of snow per unit area of the surface ($kg\ m^{-2}$).
M	Charnock-formula coefficient; Eqn (12).
M_s	Rate of snowmelt and/or icemelt ($kg\ m^{-2}s^{-1}$).
p, p_o	Atmospheric pressure; p_o is the surface pressure. mb is the unit in common use. SI unit: Pa (1 Pa = 1 Nm^{-2} = 10^{-2} mb).
P_a	Annual period i.e. 1 year (s).
P_r, $P_{r\ell}$	Surface rainfall rate: P_r is the grid-box average value; $P_{r\ell}$ is a 'local' rainfall intensity ($kg\ m^{-2}s^{-1}$).
P_s	Surface snowfall rate ($kg\ m^{-2}\ s^{-1}$).
q, q_o, $q_{sat}(T_o)$	Mean specific humidity: q_o is the surface specific humidity; $q_{sat}(T_o)$ is the saturated specific humidity corresponding to a surface temperature, T_o.
Q, Q_f	$Q = L_iE$ represents the latent heat flux due to surface evaporation or sublimation; Q_f represents the latent heat flux required to effect phase changes associated with melting or freezing at the surface (Wm^{-2}).
r_a, r_c, r, r_g	r_a is the aerodynamic resistance; r_c is the canopy or surface resistance; r is a surface-resistance term used in the Meteorological Office 11-layer AGCM; r_g is a surface-resistance term used in the Meteorological Office mesoscale model. These resistances arise in the formulae used to determine surface evaporation rates (sm^{-1}).
r_i	Soil-depth ratio, $r_i = \delta_i/\delta_o$ ($i = 1, 2, 3$), used in multi-layer soil-temperature models.
R	Cloud reflectivity. See Eqn (3) and Figure 12.

R_N, R_{SN}, R_{LN}, $R_{S\downarrow}$, $R_{L\downarrow}$	Surface radiative fluxes: R_N is the net flux; R_{SN} is the net solar flux; R_{LN} is the net long-wave flux; $R_{S\downarrow}$ is the downward component of the solar flux; $R_{L\downarrow}$ is the downward component of the long-wave flux (Wm^{-2}).
Ri_B	Bulk Richardson number; Eqn. (11).
t	Time (s).
T_o, T	T_o is the surface temperature; T is a representative air temperature in the lowest layer of a model (K).
T_g, T_{go}, T_{gi}	Soil temperature: $T_{go} \equiv T_o$ is the surface soil temperature; T_{gi} ($i = 1, 2, 3$) is the representative soil temperature at some specified depth in the soil (K).
u_*	Surface friction velocity: $u_* = [\lvert\underset{\sim}{\tau}\rvert/\rho]^{1/2}$ (ms^{-1}).
$\underset{\sim}{V}$, V	Wind vector, $\underset{\sim}{V}$, and windspeed, $V = \lvert\underset{\sim}{V}\rvert$ (ms^{-1}).
w, w'	Mean vertical component, w, of the wind vector, $\underset{\sim}{V}$. w' is the eddy fluctuation of w (ms^{-1}).
Y, Y_o, Y_δ	Y is the intensity of total runoff; Y_o is the surface runoff component of Y; Y_δ is a gravitational drainage component of Y. $Y = Y_o + Y_\delta$ ($kg\ m^{-2}\ s^{-1}$).
$Y_{o\ell}$	$Y_{o\ell}$ is a 'local' surface runoff ($kg\ m^{-2}s^{-1}$).
z	Height above the surface (m).
z_ℓ	Height of lowest model level (m).
z_o, z_ξ	z_o is the surface roughness length for momentum transfer; z_ξ is the characteristic surface roughness length for the turbulent transfer of the quantity ξ (m).
α, α_L, α_S, α_I, α_{eff}	Surface albedo: α_L is a snow-free land albedo; α_S is a snow-covered land albedo; α_I is an ice-covered land albedo; α_{eff} is an effective surface albedo (see Eqn (3)).
β, β_o	Soil moisture availability function: β_o is a particular specification of β defined in Eqn (35).
γ	Thermal inertia (or thermal property) of soil: $\gamma = (\lambda C)^{1/2}$ ($J\ m^{-2}\ K^{-1}\ s^{-1/2}$).
δ, δ_o, δ_i, δ_a	Soil depths: δ is the implied depth of the root zone; δ_o is the depth of the surface soil layer in a multi-layer soil-temperature model; δ_i ($i = 1, 2, 3$) are the corresponding depths of deeper soil layers; δ_a is the e-folding depth of the annual temperature wave in the soil (m).

Δ	Used to denote difference of two values. For example: a. $\Delta\xi(z_\ell) = \xi(z_\ell) - \xi_o$; Eqn (4), b. Δd_{sw}, change in d_{sw} over a period, Eqn (48), c. Δt, model time-step; Eqn (48).
ε	Surface long-wave emissivity or absorptivity.
ζ	Vertical co-ordinate in the soil: $\zeta = -z$ (m).
θ, θ_o	Potential temperature: θ_o is the surface value (K).
κ	Thermal diffusivity of the soil: $\kappa = \lambda/C$ ($m^2 s^{-1}$).
λ	Thermal conductivity of the soil ($Wm^{-1}K^{-1}$).
μ	Fraction of a grid-area; Eqn (43).
ξ, ξ_o, ξ_*, ξ'	Used to denote the mean value of some atmospheric variable (such as wind, temperature or specific humidity): ξ_o is the surface value of ξ; ξ_* is a scaling value for ξ in the surface layer of the atmospheric boundary layer, defined by $\xi_* = -F_\xi/u_*$; ξ' denotes the eddy fluctuation of ξ.
ρ, ρ_g, ρ_w, ρ_s	ρ is air density; ρ_g is soil density; ρ_w is liquid water density; ρ_s is snow density (kg m^{-3}).
σ	a. Vertical sigma-co-ordinate: $\sigma = p/p_o$. b. Stefan-Boltzmann constant (in Eqn (1) only) (5.670×10^{-8} Wm^{-2} K^{-4}).
$\underline{\tau}$	Surface shearing stress or momentum flux (Nm^{-2}).
ϕ_ξ, ϕ_M, ϕ_H, ϕ_E	ϕ_ξ is the Monin-Obukhov similarity function for the vertical profile of quantity ξ in the surface layer; ϕ_M, ϕ_H and ϕ_E are the appropriate functions for the turbulent transfers of momentum, sensible heat and water vapour, respectively.
χ, χ_w, χ_f, χ_c	χ is the volumetric soil water concentration: χ_w is a specified wilting point; χ_f is a nominal field capacity; χ_c is a critical value of χ below which $\beta<1$ (see Eqn (37)).
ω_a, ω_o	Angular frequency of oscillation: $\omega_a = 2\pi/P_a$ corresponds to the period P_a of 1 yr; ω_o is a characteristic frequency, determined empirically, which is used to determine δ_o, the thickness of the surface soil layer in the multi-layer soil-temperature scheme discussed in Section 5 ($\delta_o=(2\lambda/C\omega_o)^{1/2}$; Eqn (28)) ($s^{-1}$).

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By D J Carson

1. Introduction

The atmospheric boundary layer is the lowest layer of the atmosphere characterized by significant vertical flux divergences of momentum, heat and moisture which result directly or indirectly from interactions between the atmosphere and the underlying surface. It is the turbulent nature of boundary-layer flows which is vital for the efficient exchange of momentum, heat and moisture between the earth's surface below and the 'free' atmosphere above. In general, up until fairly recently designers and users of global atmospheric general circulation models (AGCMs) and operational numerical weather prediction (NWP) models have not been concerned with the details of boundary-layer and surface properties and processes in their own right but mainly for the influence they exert on weather systems and circulation characteristics on the much larger, synoptic or even global scales. Pressure to change this attitude has come from two largely disparate directions, viz:

a. following the inception of the World Climate Programme in 1979, the upsurge in development of three-dimensional AGCMs for the study of climate and climatic change and, in particular, the high priority given to the study of land-surface processes and the need to simulate near-surface climatologies;

b. the introduction into operational service of increasingly sophisticated and more highly resolved NWP models, including mesoscale models with relatively very fine resolution for local forecasting, which, in principle, should be capable of resolving some of the vertical structure in the boundary layer and are expected to forecast the near-surface meteorological variables, and even changes in the surface properties.

More interest then is being directed now towards delineating detail in the boundary layer and the characteristics of near-surface climatologies. These 'operational' needs, supported by increasingly more powerful computing systems, have led to an apparent convergence of the approaches to physical parametrization adopted by both the NWP modeller and the climate modeller. An organisation like the Meteorological Office, which is involved actively in the full range of operational forecasting and climate modelling activities, can capitalise on this common goal by concentrating appropriate expertise in a single location and by sensible unification of the relevant components of the various models to enable ready transfer of computer programs, thus avoiding unnecessary duplication of effort.

The degree to which such models can, in the longer term, be unified in the context of their representation of boundary-layer and surface processes is a moot point. In a somewhat subjective review, Smith and Carson (1977) produced a table to help determine those features of the boundary layer and the surface that need to be represented in numerical models. Having identified three classes of model as short-range (0-2 days), medium-range

(3-12 days) and long-range or climatological (> 2 weeks) and the different spatial and temporal averaging scales associated with each, their next step was to list what they called the models' 'essential practical properties', ie the products that each model was expected to simulate well. Only then did they attempt to define the boundary-layer and surface properties it would be necessary to model in order to achieve each model's practical ends. Although, as noted above, distinctions in the needs of the various models are less obvious now than Smith and Carson suggested, nevertheless their general philosophy still seems valid. Thus, in the context of boundary-layer, surface and sub-surface modelling, the ultimate needs of a global numerical model, integrated forward for perhaps several or more decades to simulate the fully interactive climate system with its very wide range of characteristic response times, are likely to differ somewhat from those of the corresponding model being used for short-range weather forecasting.

The main aim of this paper is to indicate the variety of sub grid-scale physical interactions between the atmosphere and the underlying land surface which must be represented by 'parametrizations' in current NWP and climate models. These 'land-surface processes' are delineated by describing briefly the particular parametrizations currently in use in four Meteorological Office, three-dimensional, primitive-equation models, viz:

a. the 11-layer Atmospheric General Circulation Model used for climate-modelling studies, denoted henceforth as AGCM;

b. the Global (coarse-mesh) version of the operational numerical Weather Prediction Model, denoted GWPM;

c. the Regional (fine-mesh) version of the operational numerical Weather Prediction Model, denoted RWPM;

d. the Mesoscale numerical Weather Prediction Model currently undergoing operational local forecasting trials, denoted MWPM.

Carson (1982a) provided a fairly comprehensive review of the then current practices relating to the parametrization of land-surface processes in the main three-dimensional, atmospheric general circulation models being developed as climate models at a variety of modelling centres throughout the world. The framework of this paper is similar, but in general the physical basis of a parametrization is not discussed as fully nor as critically as in Carson (1982a). This paper supplements the earlier one in two respects: (1) it updates the description of the parametrizations used in the Meteorological Office 11-layer AGCM; (2) it provides a 'domestic' review of current parametrizations of land-surface processes in the Meteorological Office's main numerical prediction models. Comments below relate only to the version of each model which is either operational, in the case of the weather prediction models, or is considered the current standard or 'control' model in the case of the AGCM.

It should be noted that the parametrization of land-surface processes is a very active field of research and model development and methods labelled 'current' may quickly become superseded. New approaches ('schemes') are being developed and tested continuously. Also, the parametrizations in the Meteorological Office models should not necessarily

be regarded as 'state-of-the-art' schemes. Corresponding models elsewhere may well include more complex and/or physically more plausible schemes, and indeed viable new techniques are being pursued actively in-house. This is particularly true in the context of the extension of climate models to include yet more components of the full, highly interactive climate system which comprises atmosphere, ocean, cryosphere, land and biosphere. A single paper cannot do justice to the range and complexity of tried schemes and unresolved problems in this field; even if restricted to land-surface processes only. Recent developments in the Meteorological Office models are highlighted, deficiencies are indicated where known, and some topics deserving of further immediate research are identified.

Finally, it should be stressed that there are many modelled physical processes not considered here which have a direct and immediate effect on the surface and sub-surface but are not governed directly by, nor specified in terms of, the land-surface processes (eg radiation; rainfall and snowfall; turbulent mixing throughout the whole boundary layer; deep convection; etc). There are also other important surface factors beyond the compass of this paper: these include considerations of ocean and sea-ice surfaces and the modelling of specific orographic influences at various spatial scales. In particular, there is no discussion of the parametrization of orographic gravity-wave drag introduced recently into the AGCM and the GWPM (see Palmer et al (1985)).

2. Relevant Design Characteristics of Meteorological Office Models

It may prove instructive at times to consider the representations of the sub grid-scale land-surface physical properties and processes in the four models, identified in Section 1, in the context of how they are used and also their spatial domains and spatial and temporal resolutions. Some potentially relevant characteristics of each model are listed in Table 1, along with the most pertinent recent references. Indispensable references are the respective model-users' 'handbooks' which should be consulted for more complete and detailed information. For the AGCM see Slingo (1985); for the GWPM and the RWPM see Dickinson (1985); and for the MWPM see Golding (1986) and other contributions to the Mesoscale Documentation series.

2.1 Model integration periods

a. AGCM: The 11-layer AGCM is used principally in climate modelling research and for so-called 'annual-cycle' integrations of up to about a decade of simulated time. A version of the model is also now being used for research into the use of dynamical models in long-range forecasting on time scales of about ten days to one month ahead. The version discussed here is the control, 'fourth annual-cycle' version.

b. GWPM: The 15-level, 'coarse-mesh', operational weather prediction model is used to provide global forecasts for up to six days ahead.

c. RWPM: The 15-level, 'fine-mesh', operational weather prediction model is used principally to provide weather forecasts for the neighbourhood of the British Isles for up to thirty-six hours ahead.

d. MWPM: The 16-level, mesoscale model is being developed to provide guidance to forecasters on the local variations of weather for up to a day ahead.

2.2 Spatial resolutions

A choice of horizontal resolutions is available for the AGCM. The 2.5° x 3.75°, regular latitude-longitude resolution cited in Table 1 is that most commonly used at present and is illustrated in Figure 1. The corresponding horizontal domains and resolutions for the RWPM and the MWPM are shown in Figures 2 and 3, respectively.

Vertical resolutions in the first 3 km above the surface (i.e. in the neighbourhood of the atmospheric boundary layer) are illustrated schematically in Figure 4, where, in order to relate the models' σ-co-ordinates to height z (and vice-versa in the case of the MWPM) the standard ICAO atmosphere has been assumed. Note that parametrized turbulent mixing is not allowed to extend through more than the lowest three layers in the AGCM (i.e. effectively no further than σ=0.79 ($\approx$ 2 km)) nor through more than the lowest four layers in the GWPM and the RWPM (i.e. effectively no further than σ=0.827 ($\approx$ 1.6 km)). In contrast, in the MWPM turbulent mixing in the vertical is not constrained so formally. However, for certain calculations in the MWPM a maximum boundary-layer depth of 2 km is implied. In practice, for surface-layer calculations (see Section 4) the lowest σ-level in the AGCM (i.e. σ=0.98744) is assumed to correspond to z=100 m and similarly in the GWPM and RWPM the lowest level (i.e. σ=0.997) is assumed to be z=30 m.

2.3 Model orography

a. AGCM: The land-surface elevations in each grid-box were taken as the mean of the values from the 1° x 1° data set of Gates and Nelson (1973) (see Figure 5).

b. GWPM: The global orography used in the coarse-mesh model has been derived by interpolation from a data set of 10' grid-box mean values provided by the US Navy. See example for Eurasia in Figure 6.

c. RWPM: The fine-mesh orography is derived by first applying a simple bi-linear interpolation to the global orography. During the first three hours of the fine-mesh assimilation cycle the orography at selected European grid-points is then gradually adjusted to a grid-box mean value consistent with the RWPM's higher horizontal resolution. Figure 7 shows the orography resulting from this process.

d. MWPM: With 15 km resolution, a reasonably faithful representation of the orography of the British Isles is achieved and the coastline has a realistic shape (see Figure 3). The mountain ranges are still somewhat lower than in reality.

2.4 Types of surface

a. AGCM: Four surface types are prescribed, viz: land; sea; land-ice; sea-ice. Over land the surface prognostic variables are the temperature, the 'soil moisture content' and the snow depth.

b. GWPM/RWPM: Six surface climatic types are used, viz: arid land; temperate land; snow-covered land; land-ice; sea; sea-ice. These are modified each month to reflect the seasonally changing patterns. The surface classifications used for January and July are shown in Figures 8 and 9, respectively. At present, the only surface prognostic variable over land is the temperature.

c. MWPM: The surface categories are simply land and sea. The surface prognostic variables over land are the temperature, a nominal indicator of water lying on the surface, and a nominal snow depth.

2.5 Physical parametrizations

In addition to the land-surface processes discussed below, each model includes its particular representation of the other, standard, sub grid-scale physical and dynamical processes including, radiative and cloud processes, convection, turbulent mixing of momentum, heat and moisture throughout the boundary layer, and precipitation. The appropriate model handbook should be referred to for specific details of such parametrizations.

3. Surface Radiative Properties and Fluxes

The net radiative flux at the surface, R_N, is expressed as

$$R_N = R_{SN} + R_{LN}$$

$$= (1 - \alpha) R_{S\downarrow} + \varepsilon (R_{L\downarrow} - \sigma T_o^4) \qquad (1)$$

where R_{SN} is the net solar radiative flux (defined positive towards the surface),

R_{LN} is the net long-wave radiative flux (positive towards the surface),

$R_{S\downarrow}$ is the downward solar radiative flux,

$R_{L\downarrow}$ is the downward long-wave radiative flux,

T_o is the surface temperature,

α is the surface albedo,

ε is the long-wave emissivity or absorptivity, and

σ is the Stefan-Boltzmann constant.

The representation of $R_{S\downarrow}$ and $R_{L\downarrow}$ at the surface is an aspect of the physical parametrizations which is outside the scope of this paper. The prediction of T_o is dealt with in Section 5. The specifications of ε and α only are considered in this section.

3.1 Surface emissivity (ε)

All the models assume explicitly or imply indirectly that the surface acts like a black body for long-wave radiation with $\varepsilon=1$. The treatment of atmospheric radiation varies between the models and is a very important and active field of research and development in its own right. A fairly complex, 'fully interactive' scheme is used in both the AGCM (see Chap 4, Slingo (1985)) and in the RWPM (see Chap 9, Dickinson (1985)). However, simpler schemes are employed in both the GWPM (see Chap 8, Dickinson (1985)) and in the MWPM (see Golding (1984) or Golding and Machin (1984)). The former of these models has a 'climatological' radiation scheme based on that used in the Meteorological Office 5-layer model as described by Corby et al (1977). In such a formulation R_{LN} is represented empirically in terms of an 'effective surface emissivity' and T_o and, in this particular case, it is not immediately obvious that the true surface emissivity, ε, has been assumed explicitly or implicitly to be unity.

3.2 Surface albedo (α)

a. AGCM: At present this is the only land-surface or soil parameter which has a geographical variation in the fourth annual-cycle control version of the 11-layer AGCM. The full specification comprises:

(i) A global data set of land cover and soils data compiled at a 1° x 1°, latitude-longitude resolution by Wilson and Henderson-Sellers (1985) has been used to create a geographical distribution of the snow-free land albedo, α_L, (see Figure 10).

(ii) The albedo of snow-covered land, α_S, is expressed as a function of α_L and the surface variable snow depth,

$$\alpha_S = \alpha_L + 0.38\, d_{sw}^{1/2} \qquad (2)$$

where d_{sw} is the water equivalent depth of snow (in cm in this formula) assumed to be one-tenth of the predicted snow depth, and α_S is limited to a maximum value of 0.6. Eqn (2) is designed

to take account of the fact that as the mean snow depth increases, not only does the snow cover surface irregularities more completely but also the area of the grid-box which is snow-free is likely to decrease.

(iii) Land-ice is ascribed an albedo, $\alpha_I = 0.7$.

b. GWPM: (i) Geographical distributions of snow-free albedo used in the operational forecast models have also been derived from the 1° x 1° data set of land usage and soil type compiled by Wilson and Henderson-Sellers (1985) (see Figures 11a-d).

(ii) Over snow-covered land, $\alpha_S = 0.5$.

(iii) Over land-ice, $\alpha_I = 0.8$.

c. RWPM: The fine-mesh model values are derived by bi-linear interpolation from the global data used in the GWPM.

d. MWPM: Simply, $\alpha = 0.2$ except where it is deemed that 'water' lying on the surface exceeds 1 mm in depth and the ground is frozen. This is taken to indicate snow cover and the albedo is increased to $\alpha_S = 0.6$.

In the AGCM and the RWPM, which use similar, fully interactive radiation schemes, an 'effective surface albedo', α_{eff}, is defined which modifies the clear-sky surface albedo, α, to allow for multiple reflections of the solar radiation between clouds and the surface. Figure 12 illustrates the process envisaged which leads to the expression,

$$\alpha_{eff} = \frac{\alpha\,(1 - CR)}{1 - \alpha\,CR}, \qquad (3)$$

where C is cloud amount and R is cloud reflectivity.

4. Surface Turbulent Exchanges

4.1 The surface turbulent fluxes

It is standard practice to represent the mean vertical surface turbulent flux, F_ξ, of some conservative quantity whose mean value is ξ, by

$$F_\xi = (\overline{w'\xi'})_o = - C_\xi V(z_\ell)\Delta\xi(z_\ell) \qquad (4)$$

where $(\overline{w'\xi'})_o$ defines F_ξ in conventional turbulence notation (i.e. it is the surface eddy covariance of ξ and the vertical velocity component w),

$V(z_\ell)$ is the mean wind speed (i.e. $|\underset{\sim}{V}(z_\ell)|$) at some specified height, z_ℓ, above the surface and within the boundary layer, (in each of the models considered here, z_ℓ is regarded as the notional height of that model's first level above the underlying surface),

$\Delta\xi(z_\ell) = \xi(z_\ell)-\xi_o$ is the difference between the value of ξ at z_ℓ and its surface value ξ_o,

and C_ξ is the so-called bulk transfer coefficient, which in general is a complicated function of height, atmospheric stability, surface roughness and other physical and physiological characteristics of the surface vegetation.

The surface turbulent fluxes required to be modelled are:

a. Momentum flux ($\underset{\sim}{\tau}$) $\underset{\sim}{\tau} = \rho\, C_D\, V(z_\ell).\underset{\sim}{V}(z_\ell)$ (5)

where C_D is the traditional 'drag coefficient' and ρ is a representative mean density of the air near the surface. The horizontal shearing stress at the surface, $\underset{\sim}{\tau}$, is conventionally a vector measure of the downward flux of horizontal momentum directed parallel to the limiting wind direction as the surface is approached. In all the models discussed here it is assumed that z_ℓ is close enough to the surface that $\underset{\sim}{\tau}$ be considered parallel to $\underset{\sim}{V}(z_\ell)$. There is the additional boundary condition of NO SLIP at the surface.

b. Sensible heat flux (H) $H = -\rho\, c_p\, C_H\, V(z_\ell)(\theta(z_\ell)-\theta_o)$ (6)

where C_H is the appropriate bulk transfer coefficient, c_p is the specific heat of air at constant pressure, and the potential temperature θ is used as the temperature which is conserved in the large-scale mixing.

c. Water vapour flux (E) $E = -\rho\, C_E\, V(z_\ell)\,(q(z_\ell)-q_o)$ (7)

where C_E is the appropriate bulk transfer coefficient and q is the mean specific humidity.

To determine these fluxes the C_ξ must be prescribed or expressed in terms of modelled variables and parameters and, in addition to the variables modelled explicitly at z_ℓ, the surface temperature and humidity need to be known. The prediction of surface temperature, T_o (simply related to θ_o), is discussed in Section 5. The surface specific humidity, q_o, is not so easy to predict explicitly and its implied value is inextricably linked to the parametrization of the surface hydrology, which is discussed in Section 6. It remains here to describe how the C_ξ are modelled.

4.2 The Monin-Obukhov approach to defining C_ξ

The models dealt with here all base their specifications of C_ξ on the Monin-Obukhov similarity theory for the surface layer of the atmospheric boundary layer. It is assumed that each model's first level above the surface, at z_ℓ, is fully within the surface layer so

that the Monin-Obukhov theory is appropriate. This is certainly a very valid assumption for the MWPM (z_ℓ = 10 m) and for the GWPM and the RWPM (both with $z_\ell \approx$ 30 m) but must be approaching the limit of practicability in the AGCM which assumes a notional value of $z_\ell \approx$ 100 m (σ_ℓ = 0.98744) in its boundary-layer parametrizations.

The Monin-Obukhov similarity hypothesis (strictly for a fully turbulent surface layer under stationary and horizontally homogeneous conditions) is the most widely accepted approach for relating surface-layer gradients of wind, temperature and specific humidity to the corresponding surface turbulent fluxes such that

$$\frac{\partial \xi}{\partial z} = \frac{\xi_*}{kz}\,\phi_\xi\,(z/L) \quad . \tag{8}$$

k is the von Karman constant (0.4) and ξ_* is a scaling value for ξ in the surface layer defined by

$$\xi_* = -\frac{F_\xi}{u_*} \tag{9}$$

where u_*, the surface friction velocity, is the particular scaling value for the surface layer wind, defined by

$$u_* = \sqrt{\left(\frac{|\underline{\tau}|}{\rho}\right)} \quad . \tag{10}$$

L, the Monin-Obukhov length, is the appropriate length-scale in the surface layer and is a measure of the stability of the turbulent flow. ϕ_ξ is a universal similarity function of z/L only which may be different for each mean transferable property, ξ, and which has to be determined empirically from analysis of surface-layer data.

From Eqns (4), (7) and (8) it is relatively straightforward to show that, having specified the ϕ_ξ (z/L), then C_ξ is in general a function of z/L, z/z_o and z/z_ξ, where z_ξ is the characteristic 'surface roughness length' derived empirically from extrapolation to the surface of the vertical profiles of ξ such that $\xi(z_\xi) \equiv \xi_o$. It is convention to denote the particular surface roughness length from the wind profile as z_o: the corresponding values for temperature and humidity are defined similarly by extrapolating the empirically fitted profile functions for θ and q downward until their surface values are implied. However, it remains the common modelling practice to take the aerodynamic surface roughness length, z_o, as the appropriate z_ξ for momentum, heat and moisture transfer. On the basis of limited empirical evidence, it is also commonly assumed that $C_E \equiv C_H$, although there is slowly mounting evidence to the contrary.

It is more convenient for modelling purposes to use a bulk Richardson number for the surface layer, Ri_B, instead of z/L as the stability indicator. With all these assumptions, the C_ξ can then be expressed as functions of z/z_o and Ri_B only where

$$Ri_B = \frac{gz_\ell}{T}\,\frac{[\Delta\theta(z_\ell) + 0.61T\,\Delta q(z_\ell)]}{V^2(z_\ell)} \quad , \tag{11}$$

g is the acceleration due to gravity and T is a representative temperature for the lowest model layer. For a full description of the method and the assumptions made see, for example, Carson and Richards (1978).

The specification of z_o and the selection of the ϕ_ξ are now discussed for the particular four models under consideration here.

4.3 Surface roughness length (z_o)

4.3.1 z_o over land

a. AGCM/GWPM/RWPM: z_o = 0.1 m for all land surfaces, including snow- and ice-covered surfaces.

b. MWPM: At present this is the only one of the four models which acknowledges a geographical variation in z_o. The evaluation of an effective surface roughness length poses a variety of as yet unsatisfactorily resolved practical problems, several of which have been discussed by Carson (1986). In routine forecasting use, the values of z_o prescribed at each grid-point in the MWPM have been obtained from a map of geostrophic drag coefficients for Great Britain derived on a 10 km x 10 km grid by Smith and Carson (1977). The roughness lengths can be derived from that map by using a relationship between geostrophic drag coefficient and z_o, also given in Smith and Carson (1977). In the MWPM the actual grid-point values of z_o for full land points are constrained to lie in the range 0.1-1.0 m, the upper limit of only 1 m having been found necessary at present to suppress instabilities in the model. Over Ireland (for which no values are available from Smith and Carson (1977)) a fixed value of 0.1 m is used.

It is worth noting that Mason (1985) has challenged the use of geostrophic drag coefficients to evaluate z_o in the manner of Smith and Carson (1977), and has presented an alternative approach based on averaging drag coefficients at some specified level in the surface layer of the atmospheric boundary layer. A specific application of the two methods to a 10 km x 10 km Ordnance Survey area (OS SU96) in southern England produced a Smith-Carson value of 0.25 m and a corresponding Mason value of 0.50 m. For a fuller discussion of the pros and cons of these two approaches, and reference to other recent work on this problem, see Carson (1986).

4.3.2 z_o over sea

Although it is not the purpose of this paper to discuss in detail the models' parametrizations of 'sea-surface processes', nevertheless it would seem most pertinent to comment briefly on their sea-surface values of z_o. At present, while the sea-surface temperature is not normally treated as a prognostic variable in the models, z_o is the single most important surface parameter to be specified for sea grid-points. The importance of boundary-layer and surface parametrizations over the oceans has been emphasised recently by the international modelling community (see, for example, WCP (1983)). There is also a growing awareness of, and concern about, the deficiencies in the simulated sea-surface fluxes, in particular the momentum fluxes, which are needed with sufficient accuracy to provide realistic upper boundary forcing to (coupled) ocean models.

a. AGCM/GWPM/RWPM: z_o at sea points has a fixed single value of 10^{-4} m.

b. MWPM: z_o at sea points is related effectively to the surface wind speed, through the so-called Charnock (1955) formula:

$$z_o = \frac{M u_*^2}{g} \qquad (12)$$

where u_* (the surface friction velocity (see Eqn (10))) of the previous time-step is used and the coefficient M now has a fixed value 0.0185 following Wu (1982). Until very recently (see, for example, Golding (1984)) a value of 0.035 was used for M, the same value as used previously by Carson (1982b) to conduct sensitivity experiments with an earlier version of the AGCM but now acknowledged to be too high.

In spite of views that the Charnock formula is clearly not the full answer, such a relationship is being tried in further sensitivity studies with the AGCM and is likely to be adopted in the operational forecasting models (GWPM and RWPM).

4.4 The Monin-Obukhov similarity functions (ϕ_ξ)

The general character of the similarity functions is fairly well established over a limited range of stability conditions, centred on neutral, but their specification for more extreme stability conditions (both stable and unstable) is much more debatable and uncertain.

a. AGCM:

(i) Unstable and neutral conditions ($z/L \leq 0$)

$$\phi_H = \phi_E = \phi_M^2 = (1 - 16\ z/L)^{-1/2} \quad 0 \geq z/L \geq -1 \qquad (13)$$

(Dyer and Hicks, 1970)

(ϕ_M is the similarity function for the wind profile ie for momentum transfer).

Note that, strictly, the Dyer and Hicks (1970) formulae are limited to $|z/L| < 1$ and so for more unstable conditions another approach, based on dimensional analysis in the so-called 'free-convection limit', was invoked to give

$$C_H = C_E = A\ |\ Ri_B\ |^{1/2} \qquad (14)$$

where $$A = \{3c[(z_\ell/z_o)^{1/3} - 1]\}^{-3/2} \qquad (15)$$

and c is a constant, determined empirically as 0.84. For fuller details of the free-convection approach see Carson (1982a). The functional dependence of the C_H on Ri_B is extrapolated empirically and smoothly from the values given by the Dyer and Hicks (1970) formulae (Eqn (13)) to attain the free-convection form (Eqn (14)) for large negative values of Ri_B.

The proper behaviour of C_D in very unstable conditions is much more difficult to determine but since the surface stresses are normally quite small in highly unstable, light-wind conditions C_D is assumed simply and arbitrarily to be constant beyond some empirically determined value of Ri_B. This method leads to C_H (and C_E) > C_D in unstable conditions.

(ii) Stable conditions ($z/L > 0$)

$$\phi_H = \phi_E = \phi_M = \begin{cases} 1 + 5\ z/L & 0 < z/L \leq 1 \\ 6 & 1 < z/L < 6 \end{cases} \qquad (16)$$

(Webb, 1970)

The problem of extending the functional form of ϕ_ξ to highly stable conditions was discussed by Carson and Richards (1978). In practice the AGCM has a lower limit set for C_ξ which applies for all Ri_B greater than a certain value.

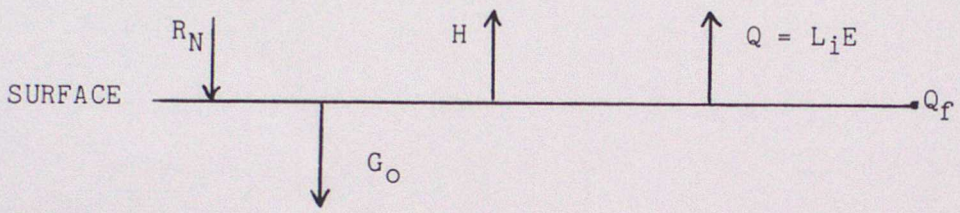
The AGCM surface-layer bulk transfer coefficients, C_D, C_H and C_E, derived from the above approach based on surface-layer similarity hypotheses are illustrated in Figure 13.

b. GWPM/RWPM/MWPM:

The same approach as detailed in 4.4a was followed in both the operational forecasting models and the mesoscale model, with minor differences only in the selections of ϕ_ξ in unstable conditions and in the subjective extrapolations needed to cater for extreme stabilities. Figure 14 illustrates the values of C_D and C_H (= C_E) used at land points in the GWPM and the RWPM. Figure 15 depicts the corresponding values used at sea points.

In the mesoscale model, MWPM, which has a grid-scale geographical distribution of z_o (see Section 4.3), the bulk transfer coefficients are calculated as functions of z_ℓ/z_o and Ri_B by bi-linear interpolation between entries in a table.

5. Land-surface Temperatures and Heat Balance



Sketch 1

Each of the models invokes an energy balance at the land surface (see above sketch) expressed most generally as

$$G_o = R_N - H - Q - Q_f \qquad (17)$$

where $Q = L_iE$ represents the latent heat flux due to surface evaporation or sublimation by turbulent transfer (L_i is the latent heat of evaporation, L_e, or sublimation, L_s),

$Q_f = L_f M_s$ represents the latent heat flux required to effect phase changes associated with melting or freezing at the surface (L_f is the latent heat of fusion and M_s is the rate of snowmelt and/or icemelt), and

G_o represents a flux of heat into the soil at the surface, and which, conventionally, is assumed to be positive when directed into the soil.

The models' formulations for G_o are based on considerations of heat conduction and conservation in the soil. In the absence of horizontal heat transfer and melting or freezing within the soil, the heat conservation equation for a normally assumed homogeneous soil reduces to

$$\frac{\partial T_g}{\partial t} = -\frac{1}{C}\frac{\partial G}{\partial \zeta} \qquad (18)$$

where Tg is the soil temperature, G is the soil heat flux, C is the volumetric heat capacity of the soil, $\zeta = -z$ is the vertical co-ordinate in the soil and t is time. Heat conduction in the soil is governed by

$$G = -\lambda \frac{\partial T_g}{\partial \zeta} \qquad (19)$$

where λ is the thermal conductivity of the soil. Eqns (18) and (19) yield the equation for conduction of heat in the soil, viz

$$\frac{\partial T_g}{\partial t} = \kappa \frac{\partial^2 T_g}{\partial \zeta^2} \qquad (20)$$

where κ is the thermal diffusivity of the soil, $\kappa = \lambda/C = \lambda/\rho_g c_g$, ρ_g is the uniform soil density and c_g is the specific heat capacity.

The four models discussed here all use Eqns (18) and (19) and the surface energy balance (Eqn (17)) to produce a prognostic equation for T_o (usually assumed equivalent to the soil surface temperature T_{go}).

a. GWPM: The simplest approach of this kind requires the introduction of an effective depth D and an effective surface thermal capacity $C_{eff} = CD = \rho_g c_g D$ such that

$$G_o = C_{eff} \frac{\partial T_o}{\partial t} \quad . \qquad (21)$$

Until recently a universal value for C_{eff} of 2.09×10^5 J m^{-2} K^{-1} (equivalent to that of a 5 cm layer of water) was used in the 11-layer AGCM. This value for C_{eff} is still used in the GWPM and changes in T_o with time are determined by substituting for G_o from Eqn (21) in Eqn (17). In addition, in that model, T_o is constrained to be $\leq$ 273.2 K over specified land-ice points.

b. AGCM/RWPM: The simple scheme represented by Eqn (21) has been replaced in the fourth annual-cycle version of the AGCM and in the fine-mesh operational forecasting model, RWPM, by a 4-layer soil model developed recently by Warrilow (1986a) and indicated schematically in Figure 16.

The surface temperature T_o is represented by the mean temperature of a very thin surface soil layer of depth δ_o. The rate of change of T_o with time is given by the soil heat flux divergence in the surface layer according to a simple finite difference form of Eqn (18), i.e.

$$\frac{\partial T_o}{\partial t} = \frac{G_o - G_1}{C\delta_o} \quad . \qquad (22)$$

G_o is the net imbalance of the terms on the RHS of Eqn (17) and G_1, the soil heat flux into the next layer down, is determined from the explicitly modelled soil temperature profile from the heat conduction Eqn (19) written simply as

$$G_1 = \frac{2\lambda(T_o - T_{g1})}{\delta_o + \delta_1} \quad . \qquad (23)$$

Therefore, from Eqns (22) and (23),

$$\frac{\partial T_o}{\partial t} = \frac{G_o}{C\delta_o} + \frac{2\kappa\,(T_{g1}-T_o)}{\delta_o(\delta_o+\delta_1)}\,, \qquad (24)$$

and the same general technique is used to provide the corresponding predictive equations for the temperatures of the other soil layers.

In the AGCM all four soil temperatures are treated as prognostic variables during integration of the model, with the boundary condition that the soil heat flux is zero at the lower soil boundary (i.e. $G_4 \equiv 0$ in Figure 16). In the RWPM only the top three soil temperatures are updated during a forecast, the bottom-layer temperature, T_{g3}, being held constant at its initialised value.

The values of the soil-layer depths and soil thermal properties used at present in the AGCM and the RWPM are listed in Table 2. Differences between the two models simply reflect the fact that Warrilow has developed the method further for the AGCM since earlier used values were adopted for use in the fine-mesh forecasting model.

Selection of the soil thermal characteristics C and λ

Warrilow's values of the basic soil thermal properties C and λ derive from his subjective choice of two other soil characteristics, viz:

(i) The e-folding depth δ_a of the annual temperature wave, defined by

$$\delta_a = \left(\frac{\kappa P_a}{\pi}\right)^{1/2} = \left(\frac{2\lambda}{C\omega_a}\right)^{1/2} \qquad (25)$$

where P_a is the period (i.e. 1 year) and $\omega_a = 2\pi/P_a$ is the angular frequency of oscillation. Warrilow (1986a) chooses 2.7 m as a typical value for δ_a.

(ii) The term $\gamma = (\lambda C)^{1/2}$, called the thermal inertia by Warrilow (1986a), but also known by several other names (see, for example, Sellers (1965)) including the term 'thermal property' which Sellers (1965) himself adopted. Warrilow now considers a value of about 1400 J m^{-2} K^{-1} $s^{-1/2}$ to be a more appropriate global value than the value of about 2000 J m^{-2} K^{-1} $s^{-1/2}$ used by him in earlier studies with the AGCM and still used operationally in the RWPM. The latter, higher value is thought to be typical of a wet, clay soil.

With δ_a and γ selected, λ and C follow from (i) and (ii) such that

$$\lambda = (\frac{\omega_a}{2})^{1/2} \gamma \delta_a \qquad (26)$$

and $$C = \frac{\gamma^2}{\lambda} \quad . \qquad (27)$$

The values so obtained were rounded off to $\lambda = 1.2\ Wm^{-1}K^{-1}$ and $C = 1.6 \times 10^6\ J\ m^{-3}\ K^{-1}$.

Selection of the soil-layer depths δ_i (i = 0(1)3)

On the basis of a study of the amplitude and phase response of multi-layer soil schemes to periodic surface temperature forcing, Warrilow (1986a) recommended the 4-layer soil model described above. The thickness of the surface soil layer he has defined as the e-folding depth of a characteristic frequency, ω_o, chosen so as to optimise the amplitude and phase responses of the scheme to a wide range of forcing frequencies encompassing both the diurnal and the annual frequencies. His current recommendation gives

$$\delta_o = (\frac{2\lambda}{C\omega_o})^{1/2} \qquad (28)$$

where $\omega_o = 3.5509 \times 10^{-4}\ s^{-1}$, corresponding to a period of about 4.8 hr. Such studies also enable him to specify the ratios $r_i = \delta_i/\delta_o$ (i = 1, 2, 3) such that each of the other layer depths δ_i (i = 1, 2, 3) is determined once δ_o has been specified.

At present only one set of soil characteristics is used in each of the AGCM and the RWPM. The only slight extension to this (see Appendix E of Dickinson (1985)) is the use of a reduced value of $\kappa = 2.66 \times 10^{-7}\ m^2\ s^{-1}$ in the RWPM for the heat conduction between the topmost soil layer and the next layer down when land points are deemed to be snow- or ice-covered. This simple ploy attempts to represent the insulating effects of snow but, curiously perhaps, it has not been seen fit to also change the corresponding thermal capacity of the surface soil layer.

In each of the models the soil temperatures are initialised by assuming a homogeneous soil and a constant sinusoidal forcing based on the annual temperature cycle. Full details for each model are given in Dickinson (1985) and Warrilow (1986a), respectively.

c. MWPM: A similar approach to that used in the AGCM, but with only two layers in the soil, is used in the MWPM. The heat conduction equations are expressed in terms of a transformed depth co-ordinate, $\zeta' = \zeta/\sqrt{\kappa}$ such that

$$\frac{\partial T_g}{\partial t} = -\frac{1}{C\sqrt{\kappa}}\frac{\partial G}{\partial \zeta'} = -\frac{1}{(\lambda C)^{1/2}}\frac{\partial G}{\partial \zeta'} = -\frac{1}{\gamma}\frac{\partial G}{\partial \zeta'} \qquad (29)$$

and

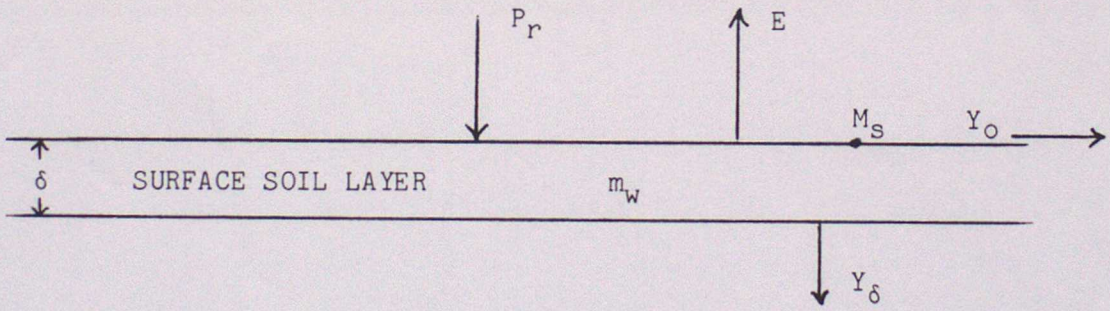
$$G = -C\sqrt{\kappa}\,\frac{\partial T_g}{\partial \zeta'} = -(\lambda C)^{1/2}\,\frac{\partial T_g}{\partial \zeta'} = -\gamma\frac{\partial T_g}{\partial \zeta'} \qquad (30)$$

where the thermal inertia, γ, has been chosen empirically as 950 J m^{-2} K^{-1} $s^{-1/2}$, a value typical of slightly moist soil (cf. the values of γ used by the AGCM and the RWPM in Table 2). The transformed depths of the two soil layers are specified so that the scheme provides a realistic response to surface temperature variations with periods of a day or less. It is not possible to infer separate values of λ and C, nor the implied actual depths of the two soil layers, from the information provided in the standard references (i.e. Carpenter (1977) and Golding (1986)).

6. Surface Hydrology and the Soil Water Budget

6.1 The soil water budget in the AGCM

Some form of surface hydrology is included in each of the models but here perhaps the variations between the models are greatest. The most complete and complicated interactive scheme is that used in the AGCM in which soil moisture and snow depth are treated as prognostic variables determined by budget equations for water, snow and heat. This model's formulation is therefore used here to illustrate the various aspects associated with the parametrization of soil hydrology.



Sketch 2

Consider the change in the soil moisture content in a single shallow surface layer of soil of depth δ. It is altered in the AGCM by rainfall, evaporation (condensation), snowmelt and runoff according to a water mass balance equation of the form:

$$\frac{\partial m_w}{\partial t} = P_r - E + M_s - Y \qquad (31)$$

where m_w is the mass of liquid water per unit lateral area in the layer considered,

P_r is the intensity of surface rainfall,

Y is the intensity of total runoff (including surface runoff, Y_o, and percolation (gravitational drainage) to lower layers, Y_δ),

and recall that E is the surface evaporation and M_s is the rate of snowmelt and/or icemelt.

m_w can be expressed in several acceptable alternative forms, for example,

$$m_w = \rho_w \delta \cdot \chi = \rho_w d_w \qquad (32)$$

where ρ_w is the density of water,
χ is the volumetric soil water concentration, and
d_w is the representative depth of water in the layer (often referred to by modellers as the 'soil moisture content').

As with the incoming radiative fluxes at the surface (see Section 3), the precipitation rate P_r (and P_s the snowfall rate) must be regarded in the present context as determined externally. It remains to describe the formulations for surface evaporation, runoff and snowmelt. A new hydrological scheme, due to Warrilow (1986b), has recently replaced the more simple and commonly adopted 'bucket' approach to runoff (see, for example, Carson (1982a)). The new scheme includes separate terms for surface runoff, Y_o, and for gravitational drainage through the surface layer, Y_δ. It is written in terms of χ and δ, the depth of the single, interactive hydrological layer of soil, is identified as the 'depth of the root zone' and is considered at present to have a universal value of 1 m.

6.1.1 Evaporation (E)

In the discussion of the surface turbulent fluxes (Section 4) it was noted that the surface specific humidity is not easy to determine. To overcome this problem the so-called 'resistance method', credited to Monteith (1965), is invoked to estimate the actual evapotranspiration from a vegetated surface, viz

$$E = \rho \frac{(q_o - q(z_\ell))}{r_a} = \rho \frac{(q_{sat}(T_o) - q_o)}{r_c} = \rho \frac{(q_{sat}(T_o) - q(z_\ell))}{r_a + r_c} \qquad (33)$$

where, from comparison with Eqn. (7), the aerodynamic resistance, $r_a = (C_E V(z_\ell))^{-1}$, represents the efficiency of the atmospheric turbulence in the evaporation process and r_c, Monteith's canopy (or surface) resistance, characterizes the physiological control of water loss by a plant community. In effect, r_c considers all the stomata of all the leaves acting in parallel, so that the canopy behaves like 'a giant leaf'. The source of water vapour in the transpiration process is the sub-stomatal cavity of the leaf where the air is saturated, or nearly so, unless the plant is under severe water stress or is dessicated. Therefore, to use this method the specific humidity in the sub-stomatal cavity is estimated as $q_{sat}(T_o)$, i.e. the saturation value corresponding to

the surface temperature, T_o, which is determined readily via the Clausius-Clapeyron relationship. The effective value of r_c for any region is a very complicated function of, for example, vegetation type and condition, surface wetness, soil moisture deficit, time of day and season of the year.

The actual formulation used for E in the AGCM is

$$E = \rho\beta \frac{(q_{sat}(T_o) - q(z_\ell))}{r_a + r} \qquad (34)$$

where $r_a = (C_E V(z_\ell))^{-1}$ as defined in Section 4. The effect of r_c in Eqn. (33) is allowed for very empirically in Eqn. (34) by the inclusion of the two terms r and β. The direct surface resistance r is simply given a single value of 60 sm^{-1} at all land points in the model. An indirect allowance for variations in r_c, particularly when the available water for evaporation is scarce (e.g. in poorly vegetated arid and semi-arid regions), is through β, an empirical 'moisture availability function' which ranges from 0 for an arid surface to 1 for a saturated surface. Even the single value of r was found to improve the surface Bowen ratio (i.e. sensible heat flux/latent heat flux) which had tended previously to be too low over land in the model.

6.1.2 The moisture availability function (β)

β has been defined for previous use in the AGCM in terms of the soil moisture content d_w such that

$$\beta \equiv \beta_o = \begin{cases} d_w/5 & 0 \leq d_w < 5 \text{ cm} \\ 1 & d_w \geq 5 \text{ cm} \end{cases} \qquad (35)$$

Warrilow (1986b) has redefined β in terms of the volumetric soil moisture concentration, χ, such that

$$\beta = \begin{cases} 0 & 0 \leq \chi < \chi_w \\ \dfrac{\chi - \chi_w}{\chi_c - \chi_w} & \chi_w \leq \chi < \chi_c \\ 1 & \chi \geq \chi_c \end{cases} \qquad (36)$$

where χ_w is called the 'wilting point', and χ_c is a critical value of the soil moisture concentration below which $\beta<1$. In addition,

$$\chi_c = \chi_w + \frac{1}{3}(\chi_f - \chi_w) \qquad (37)$$

where χ_f is a nominal value of χ for 'field capacity' and is used only to define χ_c.

The particular values of χ_w and χ_f being used in the fourth annual-cycle control version of the AGCM are listed in Table 3. In practice then Eqn (36) in the AGCM is

$$\beta = \begin{cases} 0 & 0 \leq \chi < 0.08 \\ 20\,\chi - 1.6 & 0.08 \leq \chi < 0.13 \\ 1 & \chi \geq 0.13 \end{cases} \quad (38)$$

It remains common practice in many models (see Carson (1982a)) to define a β-function to relate actual evaporation, E, to the 'potential evaporation', E_p, such that

$$E = \beta E_p = \beta \rho C_E V(z_\ell)(q_{sat}(T_o) - q(z_\ell)) \quad (39)$$

and, indeed, this is how β_o, as defined in Eqn (35), was used in earlier versions of the 11-layer AGCM. For reasons, discussed to some extent in Carson (1982a) (and more fully in Carson and Roberts (1977)), the $\Delta q(z_\ell)$ needed to evaluate Ri_B in Eqn (11), and hence the bulk transfer coefficients C_D, C_H and C_E, is still defined by

$$\Delta q(z_\ell) = \beta\,(q(z_\ell) - q_{sat}(T_o)) \quad (40)$$

which yields

$$q_o = \beta\, q_{sat}(T_o) + (1-\beta) q(z_\ell) \quad (41)$$

This would be tantamount to expressing E as,

$$E = \rho\beta \frac{(q_{sat}(T_o) - q(z_\ell))}{r_a} \quad (42)$$

from the first expression for E on the RHS of Eqn (33), which is difficult to reconcile fully with the use of the same β (as defined in Eqns (36)-(38)) in Eqn (34). Such apparent inconsistencies reflect the rather crude and empirical nature of certain areas of the parametrization of land-surface processes in existing models.

6.1.3 Runoff ($Y = Y_o + Y_\delta$)

A scheme proposed by Milly and Eagleson (1982) has been adapted for use in the AGCM. It simulates directly two components of the total runoff, viz,

(i) excess of rainfall rate over infiltration rate i.e. surface runoff, Y_o, and

(ii) gravitational drainage of existing soil water to deeper levels, Y_δ.

a. Surface runoff (Y_o)

An attempt has been made to allow for the spatial variability of rainfall since use of grid-box averages would give marked underestimation of the surface runoff. The rain is

assumed to fall over a proportion μ of the grid-box where at present μ is chosen arbitrarily as 1 for the model's so-called 'large-scale dynamic rain' and as 0.3 for its 'convective rain'. The local rainfall rate, $P_{r\ell}$, throughout a grid-area is treated statistically as represented by the probability density function

$$f(P_{r\ell}) = \frac{\mu}{P_r} \exp\left(\frac{-\mu P_{r\ell}}{P_r}\right) \qquad (43)$$

where P_r is the model's grid-point rainfall rate which is taken to represent the average grid-box rainfall.

The local surface runoff, $Y_{o\ell}$, is defined by

$$Y_{o\ell} = \begin{cases} P_{r\ell} - F & P_{r\ell} > F \\ 0 & P_{r\ell} \leq F \end{cases} \qquad (44)$$

where F is a surface infiltration rate, deemed constant for a given soil, and at present given a fixed global value (equivalent to 13 mm h^{-1}) in the fourth annual cycle control version of the AGCM. Integration of $Y_{o\ell}$ over all values of $P_{r\ell}$ yields an expression for the total surface runoff rate for a grid-area, viz,

$$Y_o = P_r \exp(-\mu F/P_r) \qquad (45)$$

b. Gravitational drainage (Y_δ)

This is represented by a hydraulic conductivity of the soil, K, prescribed as a function of χ which is assumed to be spatially homogeneous in the single soil layer. Y_δ is therefore defined by

$$Y_\delta = \rho_w K(\chi) = \rho_w K_s \left(\frac{\chi - \chi_w}{\chi_s - \chi_w}\right)^c \qquad (46)$$

where χ_s is the saturation value of χ, K_s is the saturated conductivity (i.e. $K(\chi_s)$) and c is a constant (see Table 3).

6.1.4 Budget equation for snow

A budget equation corresponding to that used for water is also used for snow on the 'surface', viz

$$\frac{\partial m_s}{\partial t} = P_s - E_s - M_s \qquad (47)$$

where m_s is the mass of snow per unit area of the 'surface', often represented as a snow depth, $d_s = m_s/\rho_s$ where ρ_s is the density of the snow, or more commonly as an equivalent depth of water $d_{sw} \approx 0.1\ d_s$. P_s is the rate of intensity of snowfall and E_s is the rate of sublimation at the surface.

When snow is lying T_o is not allowed to rise above 273 K and the snow depth accumulates without limit or decreases according to the net value of (P_s-E_s). If, however, snow is lying and the solution of the heat balance Eqn (17), excluding the term Q_f, produces an interim surface temperature value $T_o'>273$ K then sufficient snow (if available) is allowed to melt to maintain $T_o=273$ K. Melted snow is added directly to the soil moisture content with corresponding reduction in the snow depth. The heat to melt the snow and reduce T_o to 273K is effectively extracted from the surface thermal soil layer such that

$$Q_f = L_f M_s = -L_f \rho_w \frac{\Delta d_{sw}}{\Delta t} = C_{eff} \frac{(T_o'-273)}{\Delta t} \qquad (48)$$

where C_{eff} is the effective thermal capacity of the model's surface soil layer, Δt is the appropriate model time-step and Δd_{sw} is the change in water-equivalent snow depth resulting from the melting.

6.2 Soil hydrology in the GWPM and the RWPM

The representation of surface and soil hydrological processes is much more rudimentary in the operational forecasting models than in the AGCM. In fact, a surface water mass balance equation is not invoked in the GWPM and the RWPM. The evaporation from the surface is governed by Eqn (34). In the coarse-mesh model (i.e. GWPM) $r\equiv 0$, whereas in the fine-mesh model (i.e. RWPM) r is specified with a simple seasonal variation such that $r = 0$ at the beginning of the months October-May and $r = 60\ sm^{-1}$ at the beginning of the summer months June-September. Intermediate values, presumably through May and September, are calculated by linear interpolation. $r = 0$ at all times over snow and ice points.

Although β is defined in terms of the soil moisture content d_w exactly as β_o in Eqn (35), it is initialised generally very simply and held fixed throughout the integrations. In practice $\beta = 1$ when $q_{sat}(T_o) < q(z_\ell)$, irrespective of underlying surface. When $q_{sat}(T_o)>q(z_\ell)$, $\beta = 0$ over arid land points and generally speaking $\beta = 1$ over temperate land points, except possibly during very dry summer conditions when d_w might be reduced in the fine-mesh model by manual intervention. $\beta = 1$ over snow and ice points.

Snow depth is not a predicted variable in the GWPM and the RWPM.

6.3 Soil hydrology in the MWPM

The mesoscale model's surface hydrology is also relatively simple compared with that of the AGCM. However, it nevertheless has some interesting, novel features.

Evaporation is calculated according to the Monteith formulae of Eqn (33), i.e. without the additional β-term used in the other models. The surface resistance, r_c, is given a broadly realistic diurnal variation such that

$$r_c = \begin{cases} 0 & \text{if } E < 0 \text{ (i.e. dew forming)} \\ r_g & \text{if } E > 0,\ R_{s\downarrow} > 0 \text{ (day)} \\ 3r_g & \text{if } E > 0,\ R_{s\downarrow} < 0 \text{ (night)} \\ 50\ sm^{-1} & \text{if } E > 0 \text{ and surface water from rain or dew is lying.} \end{cases} \quad (49)$$

To provide r_g, MORECS soil-moisture-deficit data for real land use (i.e. on a 40 x 40 km^2 scale) are interpolated onto the mesoscale grid and the MORECS algorithm giving surface resistance for a grass surface is applied (Thompson et al, 1982). $r_g = 100\ sm^{-1}$ is used where MORECS data are not available.

The mesoscale model only monitors intercepted or surface-lying water, which accumulates due to grid-scale precipitation (and dew) and is diminished by evaporation. Runoff occurs to prevent 'surface water' exceeding 1 mm in depth and, in order to prevent the soil becoming too wet, a further reduction of 1% of surface-lying water is deemed to have infiltrated below the surface at each time-step.

The mesoscale model's treatment of snow on the surface is very crude. The surface water budget can exceed 1 mm in depth if the ground is frozen and then snow cover is diagnosed. If the surface temperature rises above freezing then all the snow is considered to have melted, the surface water in excess of 1 mm depth is run off immediately and there is no feedback on the surface energy balance equation.

7. Further Needs

It should be evident from Sections 3-6 that, in many respects, the representations of land-surface processes in the Meteorological Office 11-layer AGCM and the numerical weather prediction models (GWPM, RWPM and MWPM) are still rather crude and simple. Carson (1982a) indicates that the same is true of others' models. The demands for improvements will come from both climate modelling studies and numerical weather forecasting. Indeed, the steadily increasing number of studies with AGCM's has already amply demonstrated the sensitivity of such models to surface properties and processes (see, for example, recent reviews by Mintz (1984), Rowntree (1983, 1984) and Rowntree et al (1985)). Parametrizations thought adequate at present will undoubtedly be seen to be deficient in models which couple interactively further components of the climate system. This is already apparent with respect to air-sea interactions in coupled ocean-atmosphere models. Although the major developments in the longer term are more likely to come from climate modelling studies, nevertheless valuable feedback is being obtained from the continuous close scrutiny of the various models' performances in the acutely critical arena of operational weather forecasting - especially of local, near-surface variables such as wind and temperatures.

A schematic résumé of the processes, variables and parameters involved in the specification of land-surface parametrizations in the 11-layer AGCM is given in Table 4. The following list indicates some known important aspects of the overall problem which warrant further, immediate attention.

a. There is a need to investigate the use, particularly in the AGCM, of further geographically varying land-surface properties, based initially for example, on the soils and land-cover data sets produced by Wilson and Henderson-Sellers (1985) (see also Wilson (1984)).

b. More understanding is required of all the air-sea interactions, particularly in conditions of high surface winds, extremes of stability and generally disturbed synoptic conditions. See, for example, the discussions in Carson (1983) and other papers in WCP (1983).

c. Little attention has yet been given (in the Meteorological Office) to the representation of the physical processes associated with snow- and ice-covered land surfaces. This omission needs to be rectified.

d. Work is in progress to incorporate a thermodynamic sea-ice model into a coupled ocean-atmosphere model (see Gordon and Bottomley (1984) and Cattle (1986)). Further work needs to be done to incorporate simple ice dynamics into the model.

e. More sensitivity studies are needed to assess the performances of current parametrizations of soil heat conduction and hydrology (see, for example, Warrilow (1986 a, b, c)). Further consideration in general needs to be given to the formulation of the models' soil thermal and hydrological characteristics.

f. The more explicit effects of vegetation in the climate system need to be studied and possibly modelled. As a start the vegetation models of, for example, Deardorff (1978), Mintz et al (1983), Dickinson (1984, 1985) and Sellers et al (1985), should be appraised critically.

g. There is scope for a more thorough study of the interaction of orography with the atmospheric flow on a range of spatial scales. In particular, the inclusion in the Meteorological Office models of a parametrization of orographic gravity-wave drag due to Palmer et al (1985) needs further careful study. Comments by, for example, Carson (1986) and Mason (1985) on the parametrization of orographic influences through z_o at smaller scales also warrant further study.

h. There is a need to consider the influence on parametrizations of spatial variability of surface characteristics and atmospheric, surface and sub-surface variables within the grid-box. Warrilow (1986b) has attempted to allow for spatial variability of rainfall across a grid-area in determining surface runoff. Carson (1986) has discussed issues related to the specification of effective z_o of heterogeneous terrain. This question of determining areally-averaged derived quantities from grid-point mean atmospheric and surface

characteristics is already an acknowledged problem in estimating the turbulent fluxes at an ocean surface in order to force ocean models (see, for example, Mitchell et al (1984) and Gill (1986)).

i. The wider complex issues of the interaction of the surface with turbulent mixing through the whole of the boundary layer and with deep convection require further study. Little is understood, for example, of how roll circulations on the scale of the boundary-layer depth, which affect the structure of the boundary layer, can influence the efficiency of the surface turbulent exchanges. The structure and dynamics of cloudy and cloud-capped boundary layers also require further study with a view to improving parametrization of the surface turbulent exchanges (see, for example, WCP (1985)).

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References

Carpenter, K.M. 1977 'Surface exchanges in a mesoscale model of the atmosphere', Unpublished Met. Office Report, Met O 11 Tech. Note No. 96.

Carson, D.J. 1982a 'Current parametrizations of land-surface processes in atmospheric general circulation models', Land Surface Processes in Atmospheric Circulation Models, Editor P.S. Eagleson, 67-108, CUP, Cambridge.

Carson, D.J. 1982b 'Comments on the sensitivity of numerical simulations to different parametrizations of the boundary-layer properties and processes', Workshop on Planetary Boundary Layer Parametrization, Reading 25-27 November 1981, 119-153, ECMWF.

Carson, D.J. 1983 'The use of observational studies in the design of oceanic atmospheric boundary layer parametrizations in atmospheric circulation models', Report of the WMO/CAS Expert Meeting on Atmospheric Boundary Layer Parametrization over the Oceans for Long Range Forecasting and Climate Models, Reading 5-9 December 1983, Appendix E, 1-10, WCP-74, WCRP, ICSU/WMO.

Carson, D.J. 1986 'Issues concerning the evaluation of effective surface roughness of heterogeneous terrain', To appear in Proceedings of the Conference on Parametrization of Land-Surface Characteristics, Use of Satellite Data in Climate Models, and First Results of ISLSCP, Rome, 2-6 December 1985. Available as Unpublished Met. Office Report, DCTN 35, Met O 20.

Carson, D.J. and Richards, P.J.R. 1978 'Modelling surface turbulent fluxes in stable conditions', Boundary-layer Meteorol., 14, 67-81.

Carson, D.J. and Roberts, S.M. 1977 'Concerning the evaluation of the turbulent fluxes at unsaturated surfaces', Unpublished Met. Office Report, Met O 20 Tech. Note No. II/103.

Cattle, H. 1986 'Tests of an ice model with leads and the impact of ice boundary changes on the atmospheric circulation - A poster presentation'. Editor H. Cattle. presented at the IAMAP/IAPSO Joint Assembly, Honolulu, 5-16 August 1985. Available as Unpublished Met. Office Report, Met O 20 Internal Note No. 43.

Charnock, H. 1955 'Wind stress on a water surface', Quart. J. R. Met. Soc., 81, 639-640.

Corby, G.A., Gilchrist, A. and Rowntree, P.R. 1977 'United Kingdom Meteorological Office five-level general circulation model', Methods in Computational Physics, 17, 67-110, Academic Press, New York.

Deardorff, J.W. 1978 'Efficient prediction of ground surface temperature and moisture, with inclusion of a layer of vegetation', J. Geophys. Res., 83, 1889-1903.

Dickinson, A. 1985 'The Weather Prediction Model', Editor A. Dickinson. Unpublished Met. Office Report, Operational Numerical Weather Prediction System, Documentation Paper No. 4, December 1985.

Dickinson, R.E. 1984 'Modelling evapotranspiration for three-dimensional global climate models', Climate Processes and Climate Sensitivity, Geophys. Monograph 29, Maurice Ewing Vol. 5, 58-72, AGU.

Dickinson, R.E. 1985 'Status of the formulation of the land-surface hydrological processes in GCMs', Report of the First Session of the JSC Scientific Steering Group on Land Surface Processes and Climate, Geneva, 21-25 January 1985, Appendix B, 1-15, WCP-96, WMO/TD-No. 43, WCRP, ICSU/WMO.

Dyer, A.J. and Hicks, B.B 1970 'Flux gradient relationships in the constant flux layer', Quart. J.R. Met. Soc., 96, 715-721.

Gadd, A.J. 1985 'The 15-level weather prediction model', Meteorol. Mag., 114, 222-226.

Gates, W.L. and Nelson, A.B. 1973 'A new tabulation of the Scripps topography on a 1° global grid. Part 1: Terrain heights,' R-1276-ARPA, Rand Corporation, Santa Monica, California, USA.

Gill, A.E. 1986 'The relationship between surface wind and surface stress', Draft TOGA Discussion Paper, Private Communication.

Golding, B.W. 1984 'The Meteorological Office mesoscale model: its current status', Meteorol. Mag., 113, 288-302.

Golding, B.W. 1986 'Surface exchanges', Unpublished Met. Office Report, Met O 11 Mesoscale Documentation Paper No. 6, Version 1 January 1986.

Golding, B.W. and Machin, N.A. 1984 'The United Kingdom Meteorological Office mesoscale forecasting system', Proc. Nowcasting - II Symposium, Norrkoping, Sweden, 3-7 September 1984, ESA SP-208, 309-314.

Gordon, C. and Bottomley, M. 1984 'A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model', Unpublished Met. Office Report, DCTN 1, Met O 20.

Mason, P.J. 1985 'On the parametrization of orographic drag', To appear in Proc. Seminar on Physical Parametrization for Numerical Models, Reading, 9-13 September 1985, ECMWF. Available as Unpublished Met Office Report, Met O 14 TDN No. 177, September 1985.

Milly, P.C.D. and Eagleson, P.S. 1982 'Parametrization of moisture and heat fluxes across the land surface for use in atmospheric general circulation models', MIT Report No. 279, Cambridge, Mass.

Mintz, Y. 1984 'The sensitivity of numerically simulated climates to land-surface boundary conditions', The Global Climate, Editor J.T. Houghton, 79-105, CUP, Cambridge.

Mintz, Y., Sellers, P.J. and Willmott, C.J. 1983 'On the design of an interactive biosphere for the GLAS general circulation model', NASA Tech. Memo 84973, Laboratory for Atmos. Sci., Modelling and Simulation Facility, Goddard Space Flight Center, Greenbelt, Maryland.

Mitchell, J.F.B., Wilson, C.A. and Price, C.M. 1984 'On the specification of surface fluxes in coupled atmosphere ocean general circulation models', Unpublished Met. Office Report, DCTN 7, Met O 20.

Monteith, J.L. 1965 'Evaporation and environment', Symp. Soc. Exptl. Biol., 19, Swansea, 8-12 September 1964, 205-234.

Palmer, T.N., Shutts, G.J. and Swinbank, R. 1985 'A parametrization of orographic gravity wave drag in a high resolution general circulation model', Research Activities in Atmospheric and Oceanic Modelling, Editor G.J. Boer, Report No. 8, September 1985, 4.10-4.13, WCRP, Num. Exptn. Prog., WMO/TD-No. 63.

Rowntree, P.R. 1983 'Land surface processes in climate models - parametrization and model sensitivity', Report of the Fourth Session of the JSC, Venice, 1-8 March 1983, Appendix B, 57-75, WCRP, ICSU/WMO.

Rowntree, P.R. 1984 'Review of general circulation models as a basis for predicting the effects of vegetation change on climate', To appear in Proc. United Nations Univ. Workshop on Forests, Climate and Hydrology - Regional Impacts, Oxford, 26-30 March 1984. Available as Unpublished Met. Office Report, Met O 20 Tech. Note No. II/225.

Rowntree, P.R., Wilson, M.F. and Sangster, A.B. 1985 'Impact of land surface variations on African rainfall in general circulation models', Unpublished Met. Office Report, DCTN 30, Met O 20.

Sellers, P.J., Mintz, Y. and Sud, Y.C. 1985 'A simple biosphere model (SiB) for use within general circulation models', Unpublished Report, Laboratory for Atmospheres, NASA/Goddard Space Flight Center, Greenbelt, Maryland.

Sellers, W.D. 1965 Physical Climatology, The Univ. of Chicago Press, Chicago and London.

Slingo, A. 1985 'Handbook of the Meteorological Office 11-layer atmospheric general circulation model. Vol. 1: Model description', Editor A. Slingo, Unpublished Met. Office Report, DCTN 29, Met O 20.

Smith, F.B. and Carson, D.J. 1977 'Some thoughts on the specification of the boundary layer relevant to numerical modelling', Boundary-layer Meteorol., 12, 307-330.

Thompson, N., Barrie, I.A. and Ayles, M. 1982 'The Meteorological Office rainfall and evaporation calculation system: MORECS (July 1981)', Hydrological Memo. No. 45, Met. Office.

Warrilow, D.A. 1986a 'Modelling of land surface processes and their influence on European climate', CEC Contract Report, CL113 UK(H).

Warrilow, D.A. 1986b 'The sensitivity of the UK Meteorological Office atmospheric general circulation model to recent changes to the parametrization of hydrology', To appear in Proc. Conf. on Parametrization of Land-surface Characteristics, Use of Satellite Data in Climate Models, and First Results of ISLSCP, Rome 2-6 December 1985.

Warrilow, D.A. 1986c 'Indications of the sensitivity of European climate to land use variations using a one-dimensional model', Ibid.

WCP 1983 Report of the WMO/CAS Expert Meeting on Atmospheric Boundary Layer Parametrization over the Oceans for Long Range Forecasting and Climate Models, Reading, 5-9 December 1983, WCP-74, WCRP, ICSU/WMO.

WCP 1985 Report of the JSC/CAS Workshop on Modelling of Cloud Topped Boundary Layer, Fort Collins, Colorado, 22-26 April 1985, WCP-106, WCRP, ICSU/WMO.

Webb, E.K. 1970 'Profile relationships: the log-linear range and extension to strong stability', Quart. J.R. Met. Soc., 96, 67-90.

Wilson, M.F. 1984 'The construction and use of land surface information in a general circulation climate model', Ph. D. Thesis, Department of Geography, University of Liverpool.

Wilson, M.F. and Henderson-Sellers, A. 1985 'A global archive of land cover and soils data for use in general circulation climate models', J. Climatol., 5, 119-143.

Wu, J. 1982 'Wind-stress coefficients over sea surface from breeze to hurricane', J. Geophys. Res., 87, No. C12, 9704-9706.

MODEL	11-layer AGCM	15-level Operational Weather Prediction Model		16-level mesoscale model MWPM
		Coarse-mesh (GWPM)	Fine-mesh (RWPM)	
Type	3-dim., finite difference, primitive equation	As for AGCM	As for AGCM	As for AGCM. Also non-hydrostatic.
Horizontal Domain	Global	Global	Regional: North Atlantic and western Europe. (80°W-40°E, 30-80°N) Fig. 2	Regional: British Isles Fig. 3
Horizontal Grid	Regular lat-long 2.5° x 3.75° (72 x 96 points) Fig. 1	Regular lat-long 1.5° x 1.875° (121 x 192 points)	Regular lat-long 0.75° x 0.9375° (65 x 129 points) Fig. 2	Regular, 15 km resolution (63 x 79 points) Fig. 3
Typical mid-latitude grid-length	~ 280 km	~ 150 km	~ 75 km	15 km
Vertical co-ordinate	$\sigma = p/p_o$ p is atmospheric pressure p_o is the surface pressure	σ	σ	z: height above underlying surface
Vertical Resolution	11 unevenly spaced levels	15 unevenly spaced levels	As for GWPM	16 unevenly spaced levels
No. of levels in the boundary layer	Up to 3 Fig. 4	Up to 4 Fig. 4	As for GWPM Fig. 4	Turbulent mixing can extend throughout whole atmosphere 7 levels: $z \leq 2110$ m 5 levels: $z \leq 1010$ m Fig. 4

TABLE 1. Design characteristics of Meteorological Office Models

MODEL	11-layer AGCM	15-level Operational Weather Prediction Model		16-level mesoscale model MWPM
		Coarse-mesh (GWPM)	Fine-mesh (RWPM)	
Lowest level (z_ℓ)	σ = 0.98744 z_ℓ ~ 100 m Fig. 4	σ = 0.997 z_ℓ ~ 30 m Fig. 4	As for GWPM Fig. 4	z_ℓ = 10 m Fig. 4
Top of the boundary layer	Up to layer boundary σ = 0.79	Up to layer boundary σ = 0.827	As for GWPM	In principle it can extend to model's highest level. Nominal upper limit of 2 km used in some calculations.
Time-step	10 min	15 min	$7^1/_2$ min	1 min
References	Slingo (1985) Warrilow (1986 a, b)	Dickinson (1985) Gadd (1985)	As for GWPM	Golding (1984,1986) Golding and Machin (1984)

Table 1 (continued)

SOIL THERMAL PROPERTIES	AGCM	RWPM
Volumetric heat capacity, C ($J\ m^{-3}\ K^{-1}$)	1.6×10^6	2.4×10^6
Thermal conductivity, λ ($W\ m^{-1}K^{-1}$)	1.2	1.8
Thermal diffusivity, $\kappa = \lambda/C$ (m^2s^{-1})	7.5×10^{-7}	7.5×10^{-7}
Thermal inertia, $\gamma = (\lambda C)^{1/2}$ ($J\ m^{-2}K^{-1}s^{-1/2}$)	1386	2078
Soil-layer depths:		
δ_o (m)	0.065	0.055
δ_1 $[r_1=\delta_1/\delta_o]$	0.254 [3.91]	0.214 [3.89]
δ_2 $[r_2=\delta_2/\delta_o]$	0.913 [14.05]	0.599 [10.89]
δ_3 $[r_3=\delta_3/\delta_o]$	2.902 [44.65]	1.378 [25.05]
$\sum_{i=o}^{3} \delta_i$	4.135	2.246
$\delta_o = (\frac{2\lambda}{C\omega_o})^{1/2}$		
ω_o (s^{-1})	3.5509×10^{-4}	4.9572×10^{-4}
$P_o = 2\pi/\omega_o$ (day)	0.2048	0.1467
Thermal capacity of surface layer of soil: $C\delta_o$ ($J\ m^{-2}K^{-1}$)	1.04×10^5	1.32×10^5

Table 2. Physical properties of Warrilow's (1986a) 4-layer soil models used for soil heat conduction in the AGCM and the RWPM. For comparison, the corresponding thermal inertia, γ, for the 2-layer soil model used in the MWPM is 950 $J\ m^{-2}K^{-1}s^{-1/2}$.

Depth of the root zone	δ (m)	1
Characteristics of the volumetric soil moisture concentration: χ		
Wilting point	χ_w	0.080
'Critical' point	χ_c	0.130
Nominal field capacity	χ_f	0.230
Saturation value	χ_s	0.445
Saturated hydraulic conductivity	K_s (mmh^{-1})	13.0
Surface infiltration rate	F (SI units: kg $m^{-2}s^{-1}$)	13.0 mmh^{-1} equivalent (= F/ρ_w)
Exponent in Eqn (46)	c	6.6

Table 3. Soil hydrological characteristics used in Warrilow's (1986b) hydrological scheme in the control, fourth annual-cycle version of the 11-layer AGCM.

	Radiative	Thermal	Hydrological	Dynamical
'External forcing'	$R_{S\downarrow}$, $R_{L\downarrow}$		P_r, P_s	
Atmospheric variables		← $\theta(T)$,	q,	$\underset{\sim}{V}$ →
Surface variables	← T_o →		q_o, d_s	
Surface parameters	α, ε	← z_o →	r, β	
Surface fluxes	R_N	H $Q = L_i E$ $Q_f = L_f M_s$ G_o	E, E_s M_s Y_o	$\underset{\sim}{\tau}$
Sub-surface fluxes		G	Y_δ	
Sub-surface parameters		λ, C	δ K Significant Values of χ	
Sub-surface variables		T_g	χ	

Table 4. Schematic representation of the processes, variables and parameters involved in the specification of land-surface parametrizations in the 11-layer AGCM.

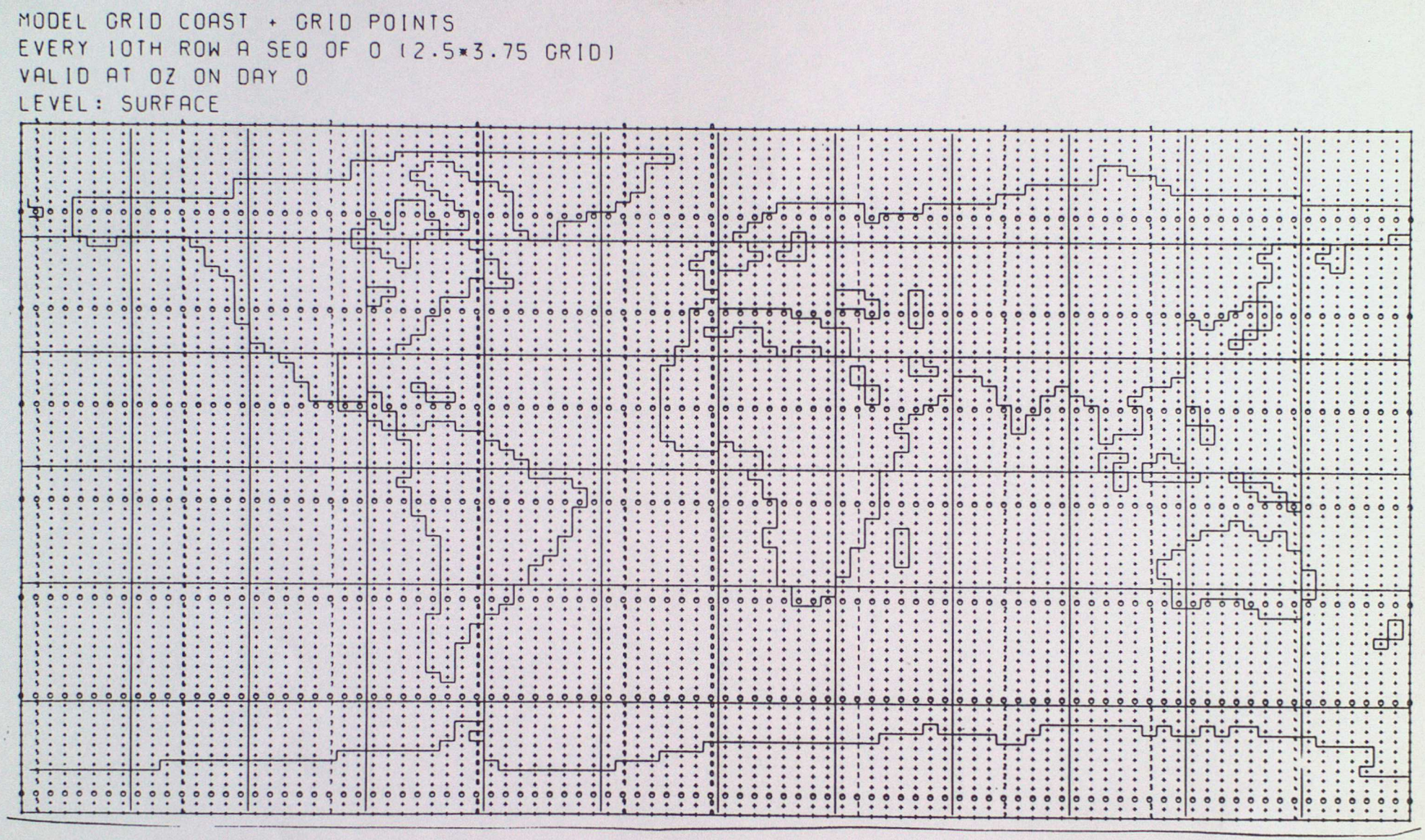


Figure 1 Horizontal resolution of the 11-Layer Model when integrated on the 2.5 x 3.75 degree latitude-longitude grid. (AGCM)

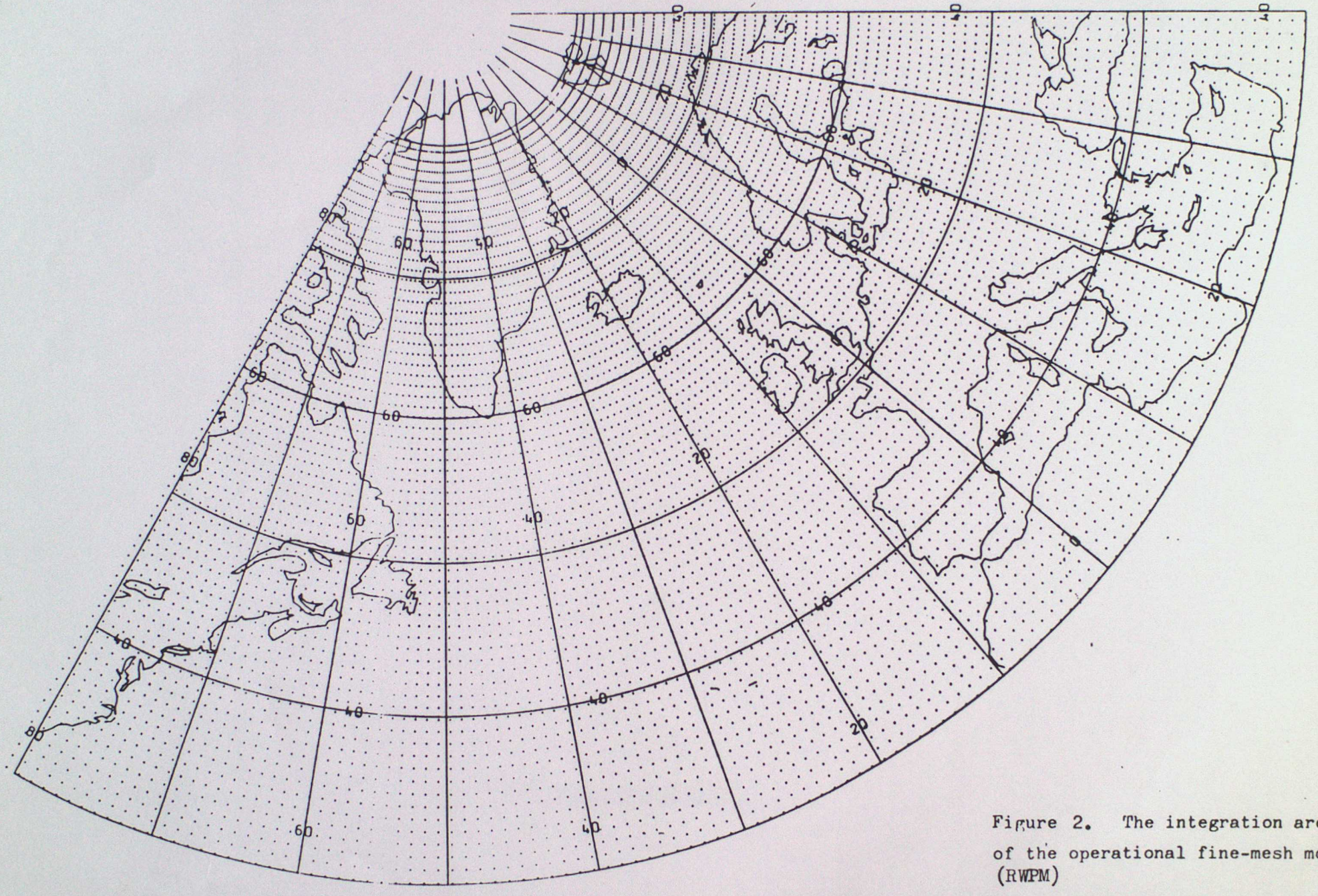
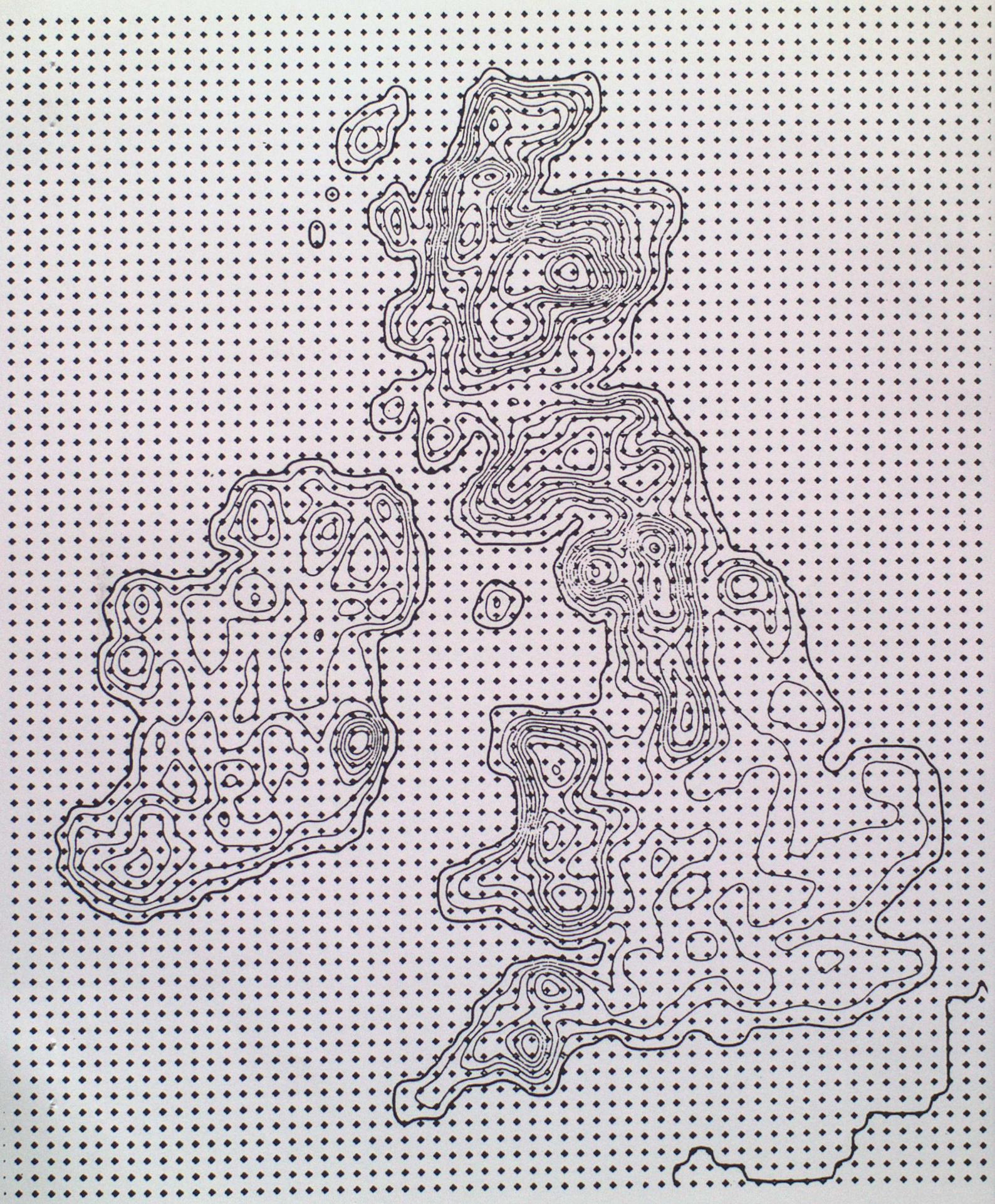


Figure 2. The integration area of the operational fine-mesh model. (RWPM)

Figure 3. The meso-scale model domain and orography. The grid points have a 15km spacing and the contour interval is 50m. The bold contours indicate the model's coastlines. (MWPM)



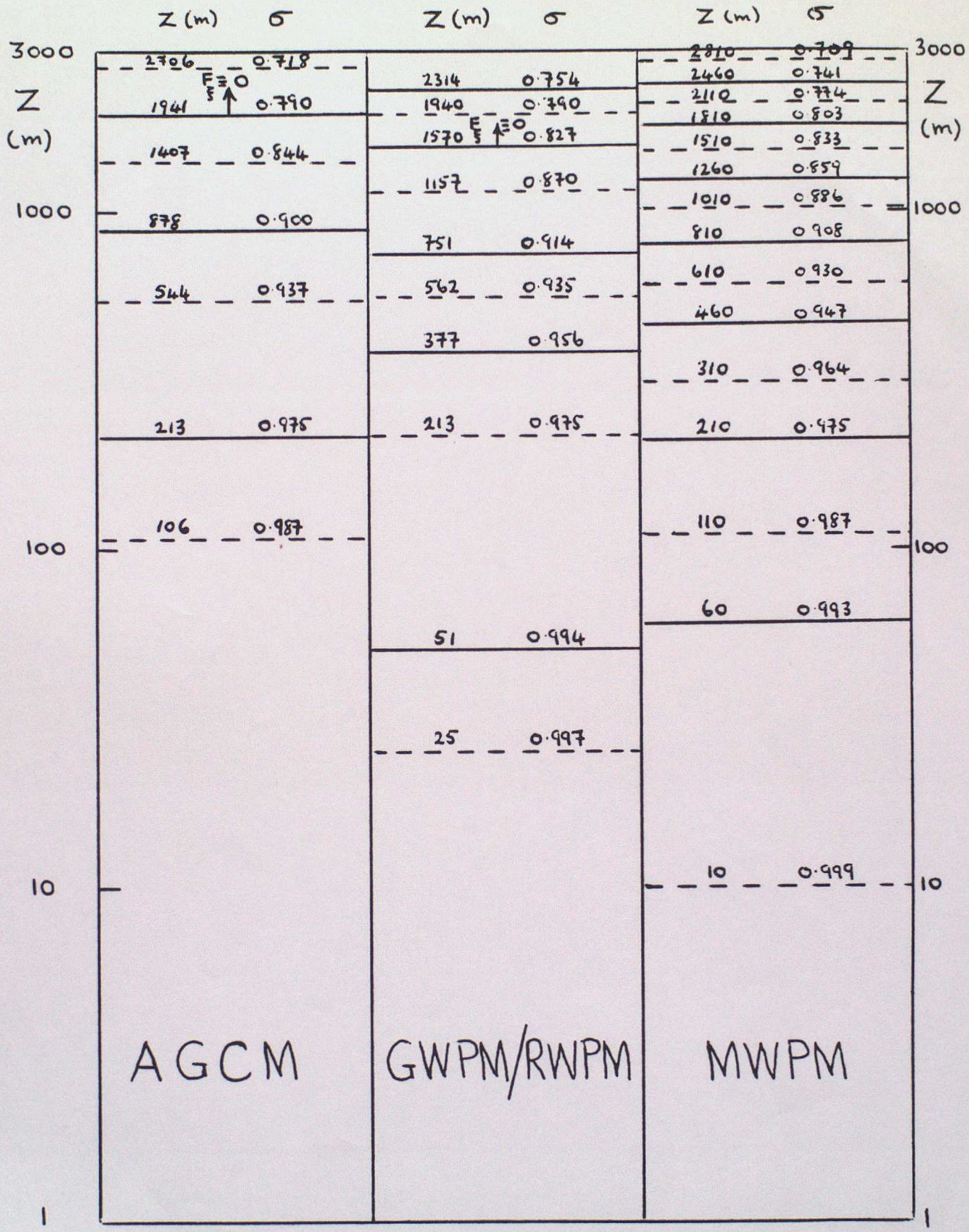


Figure 4. Schematic representation of the vertical resolution in the lowest 3 km of the Meteorological Office models. Broken lines denote model full-levels; full lines denote layer boundaries (half-levels). The vertical scale is logarithmic in height.

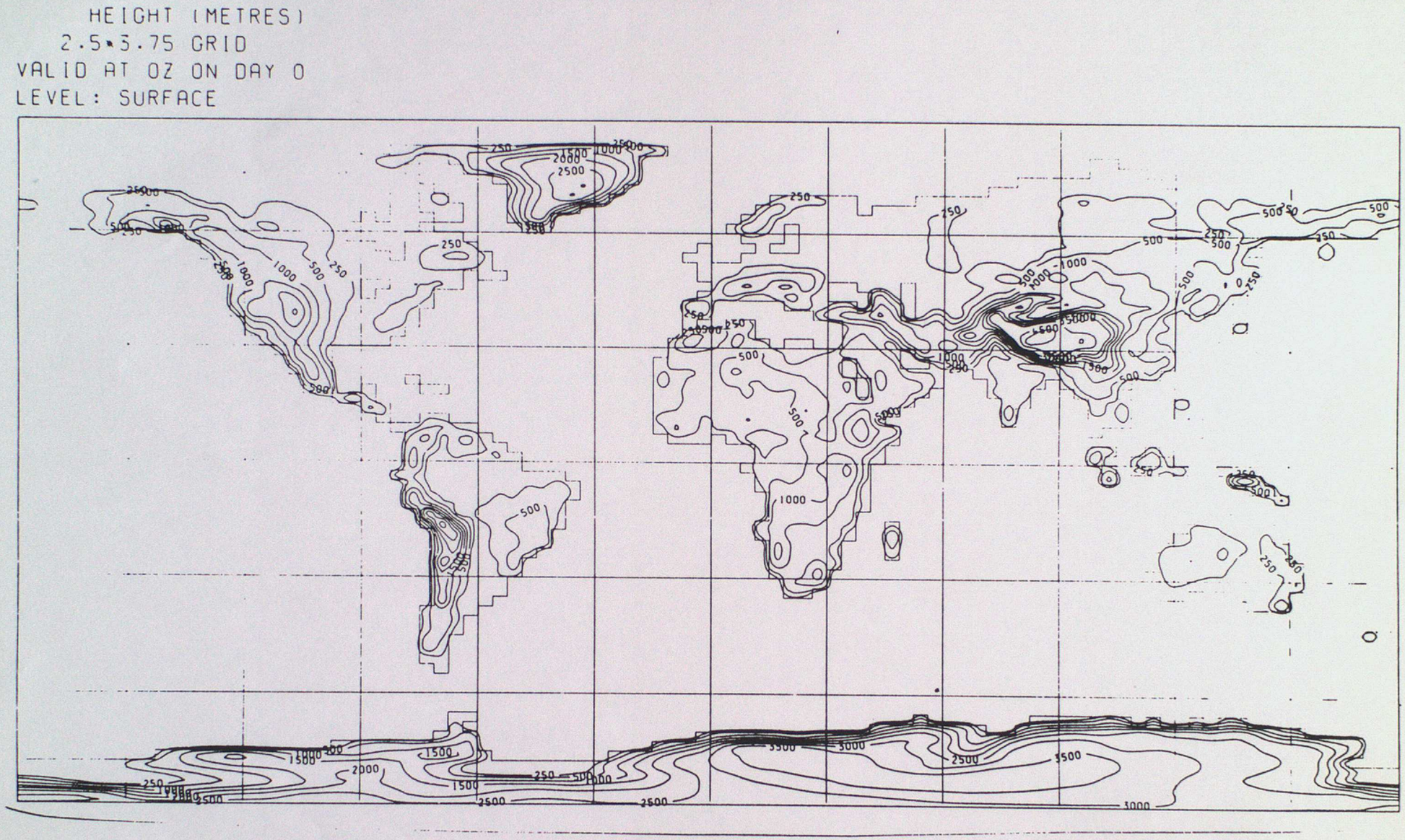


Figure 5. Surface elevations for the 2.5 x 3.75 degree latitude-longitude grid. (AGCM)

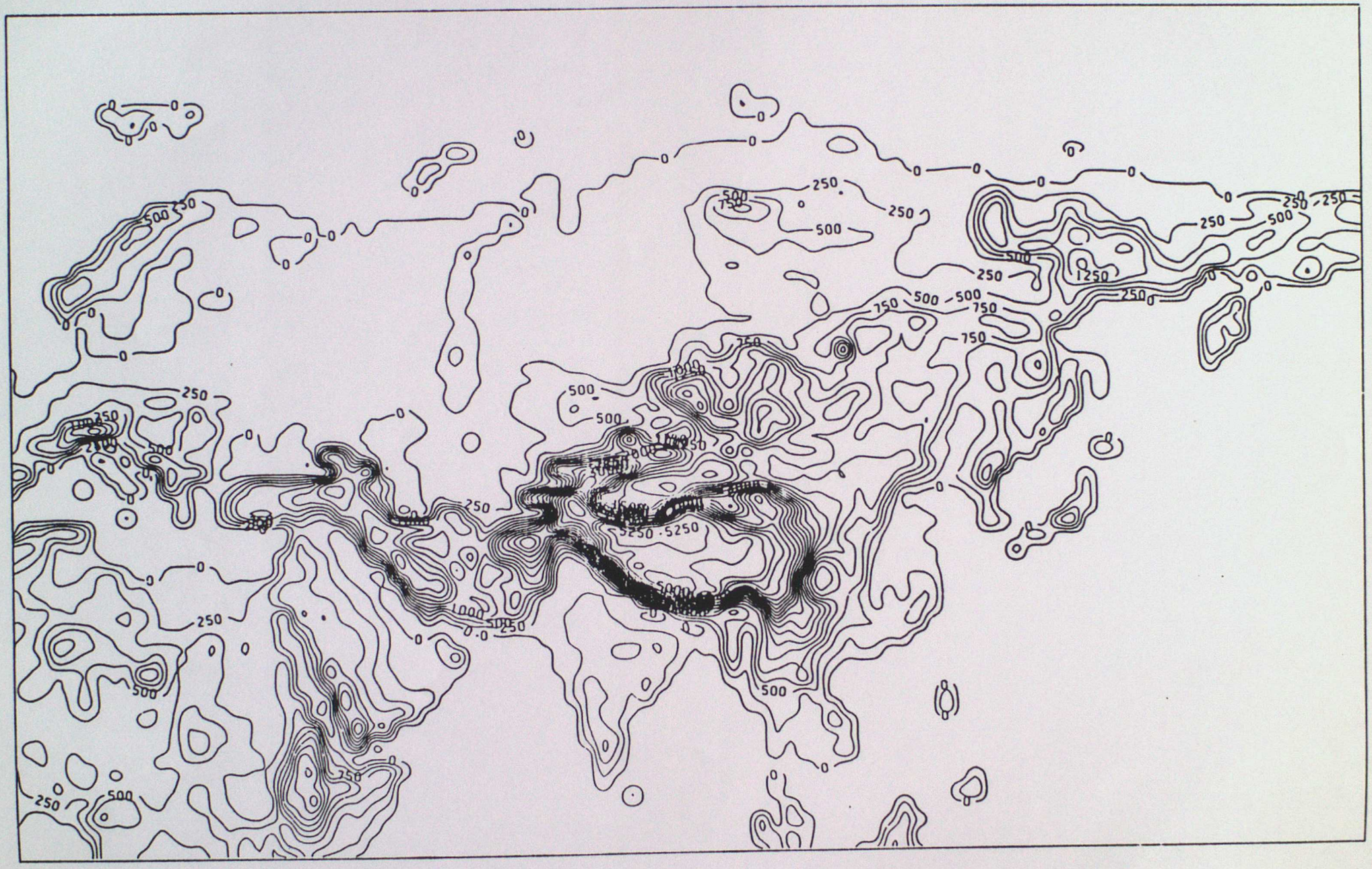


Figure 6. . Global model orography (GWPM) Contour interval = 250 m

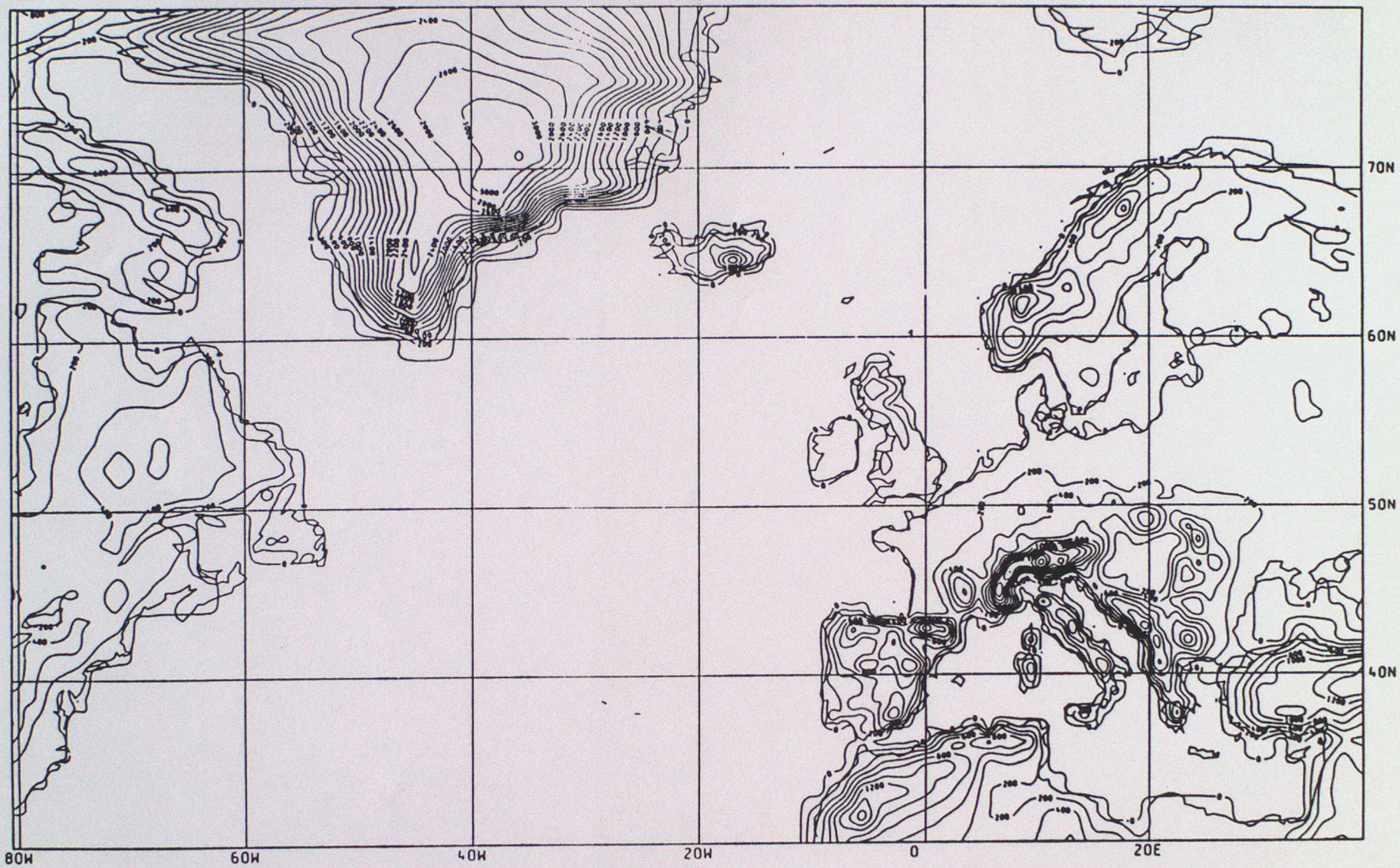


Figure 7. Fine mesh model orography. (RWPM) Contour interval = 200m

Figure 8. JANUARY SURFACE TYPES. (GWPM)

Temperate land=⊕ ;Sea-ice=⋀ ;Land-ice=△ ;Arid-land=⊡ ;Snow covered land=✳

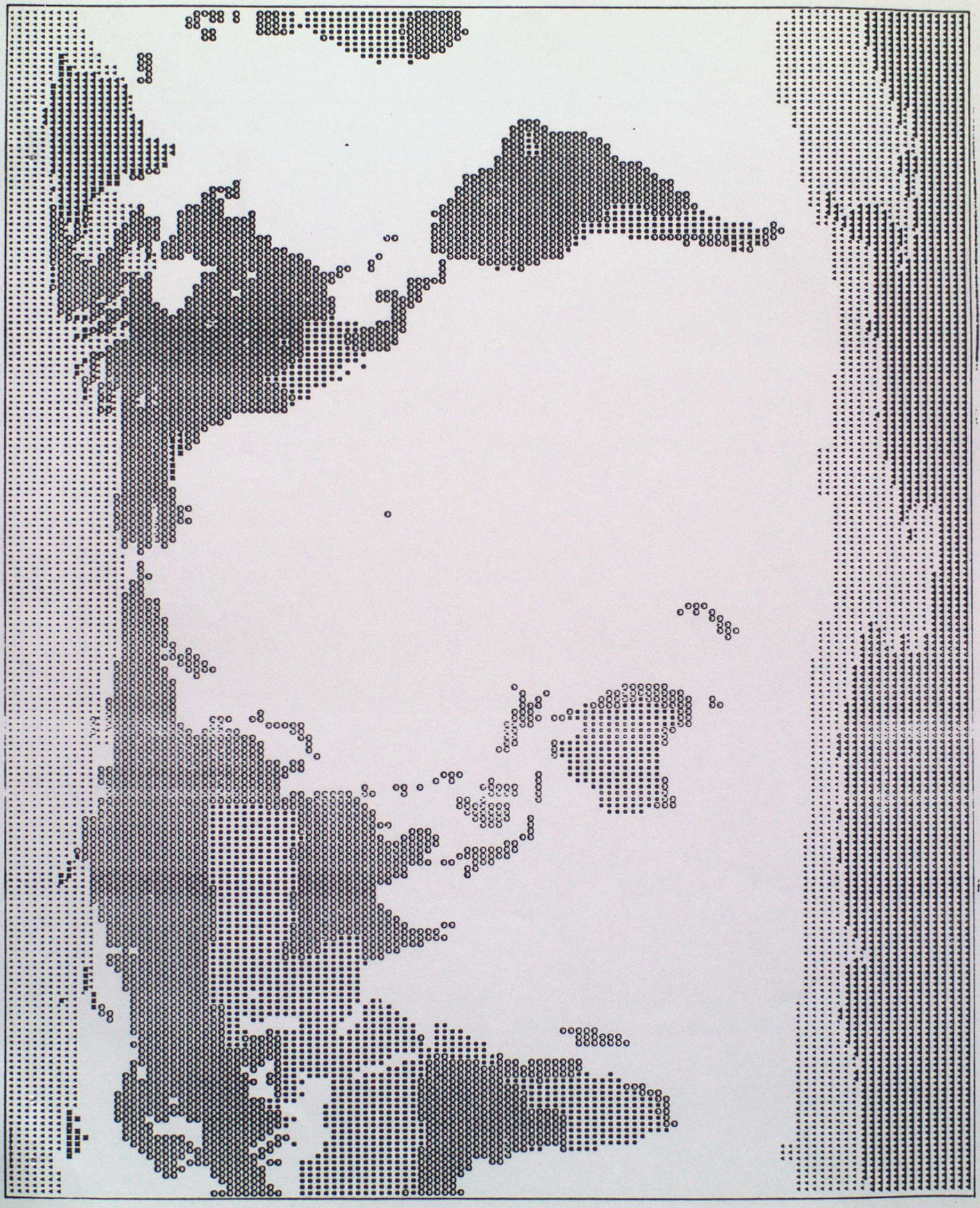


Figure 9. JULY SURFACE TYPES. (GWPM)

Temperate land=⦶ ;Sea-ice=∧;Land-ice=⚠ ;Arid-land=⊞ ;Snow covered land=✳

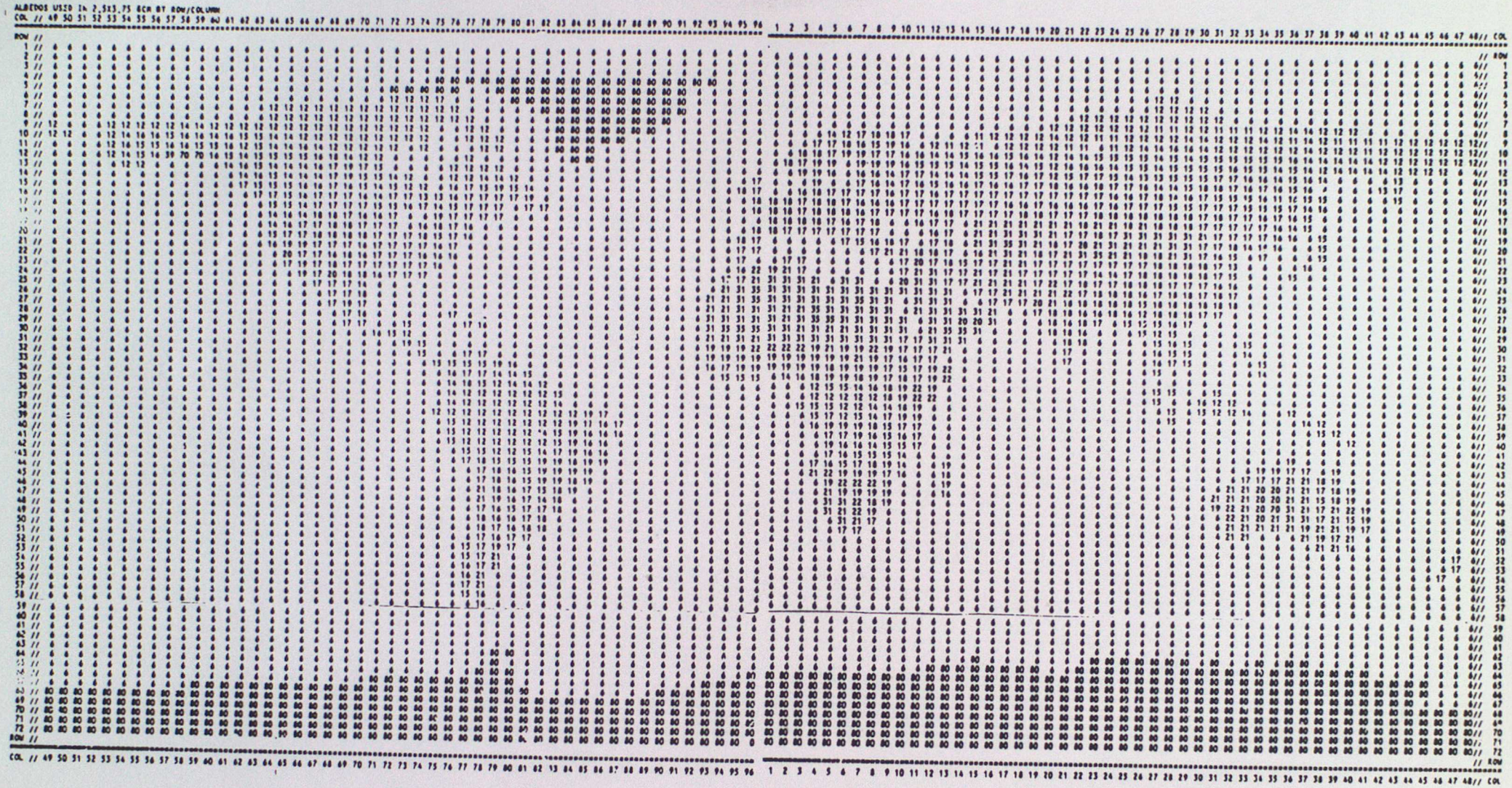


Figure 10. Snow-free surface shortwave albedos (%) used in the model when integrated on the 2.5 x 3.75 degree latitude-longitude grid. In this dataset sea-ice is ignored, so all sea points are allocated the value 6%. (AGCM)

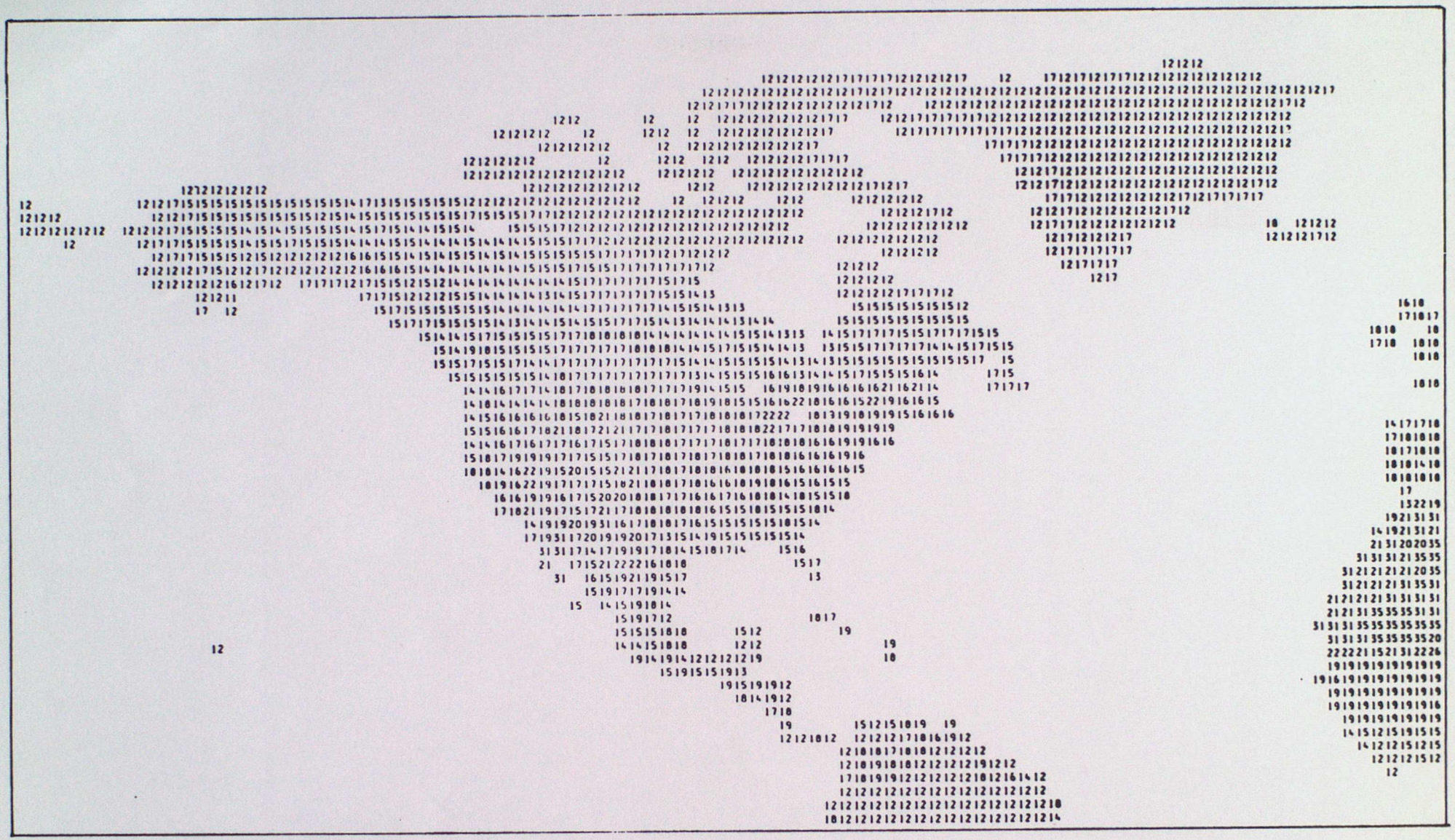


Figure 11a. Snow-free surface albedos (%) used in the global model. (GWPM)

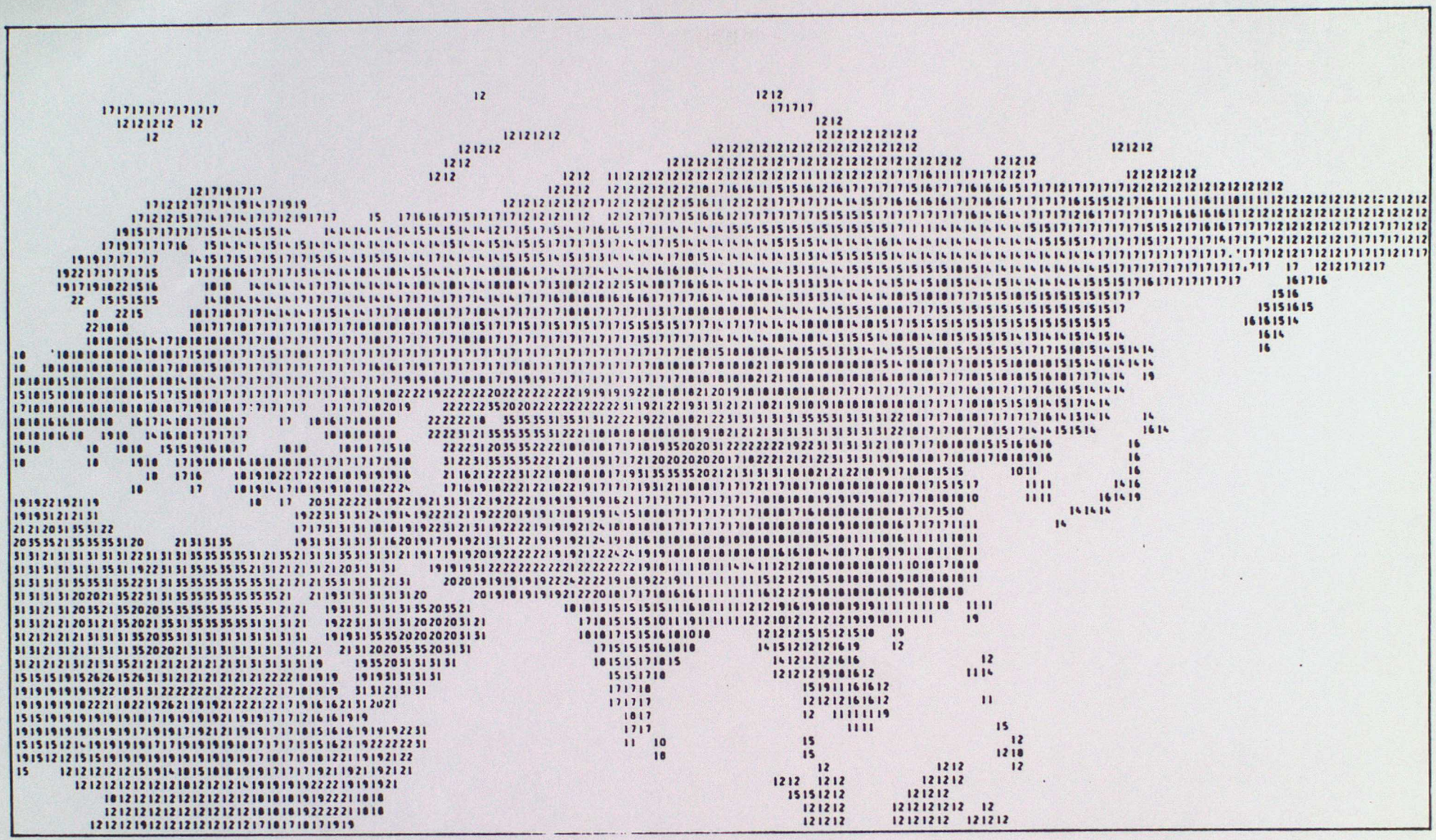


Figure 11b. Snow-free surface albedos (%) used in the global model (continued). (GWPM)

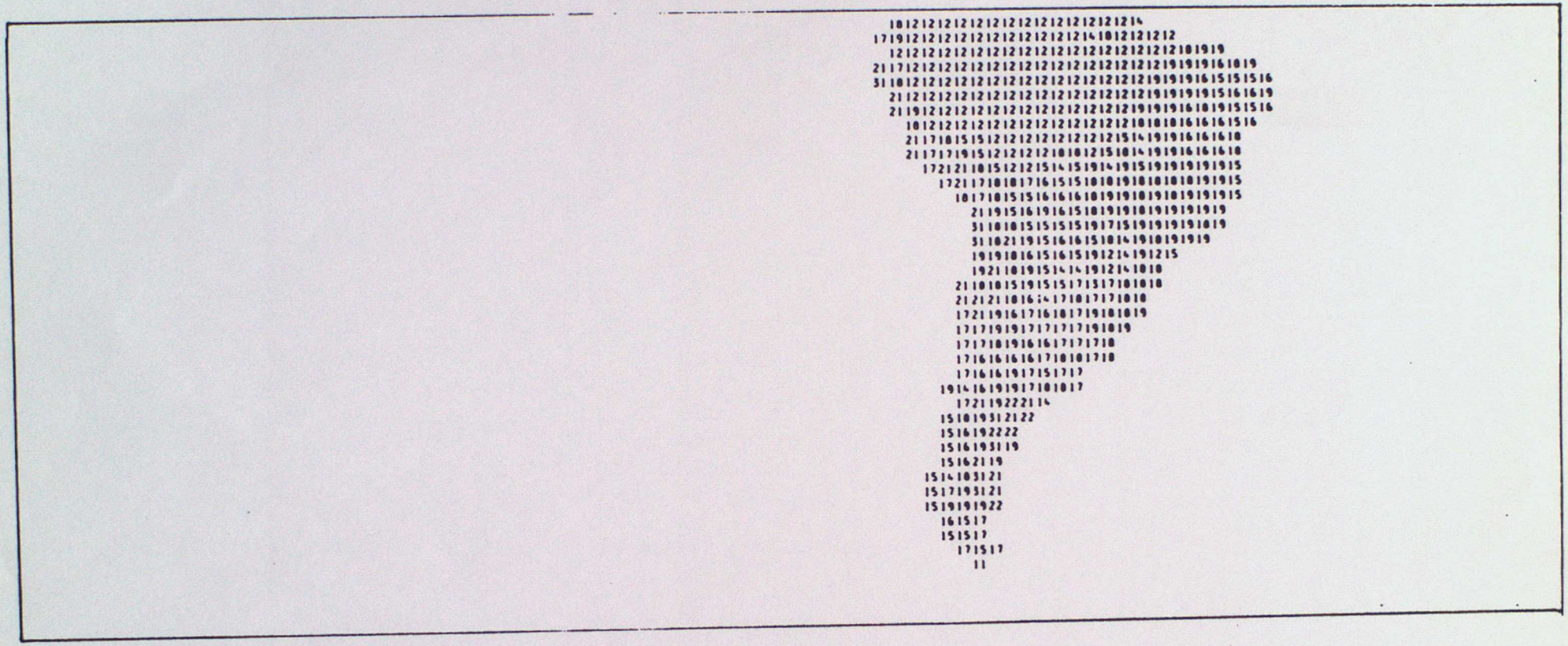


Figure 11c. Snow-free surface albedos (%) used in the global model (continued). (GWPM)

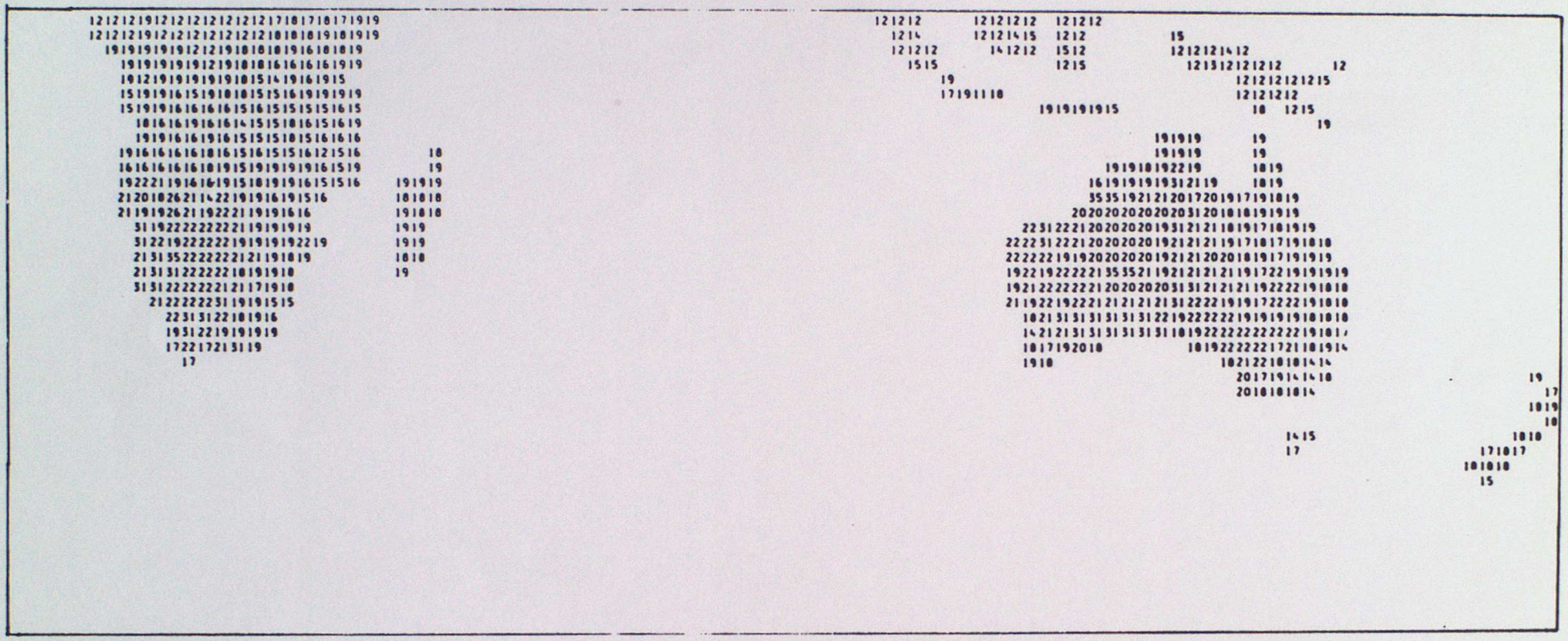


Figure 11d. Snow-free surface albedos (%) used in the global model (continued). (GWPM)

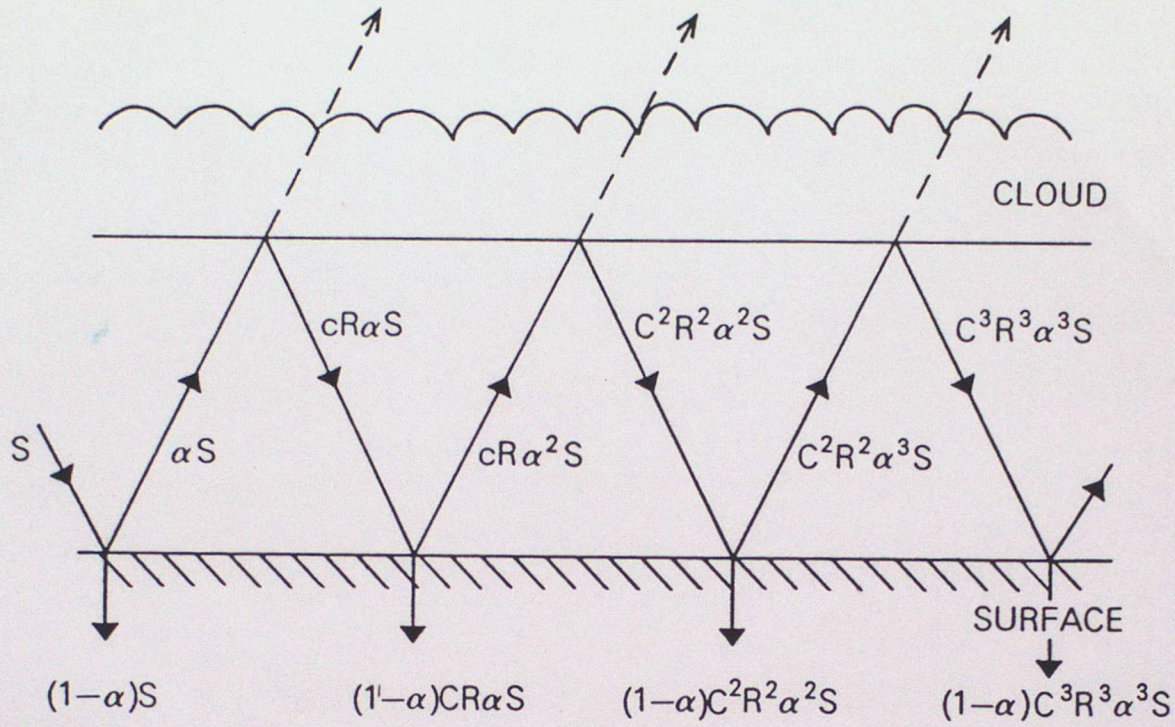


Figure 12. Multiple reflections between clouds and the surface. (AGCM and RWPM)

Let S = incoming solar flux

C = cloud amount

R = cloud reflectivity

α = surface albedo

α_{eff} = effective surface albedo

Net solar flux absorbed by surface = $(1 - \alpha_{eff})S$

$= (1-\alpha)S + (1-\alpha)\,CR\alpha S + (1-\alpha)C^2R^2\alpha^2S + \ldots.$

Hence $(1-\alpha_{eff}) = (1-\alpha)\,(1 + CR\alpha + C^2R^2\alpha^2 + \ldots.)$

$$= \frac{(1-\alpha)}{(1 - CR\alpha)}$$

Thus $\alpha_{eff} = \dfrac{\alpha\,(1 - CR)}{(1 - \alpha CR)}$

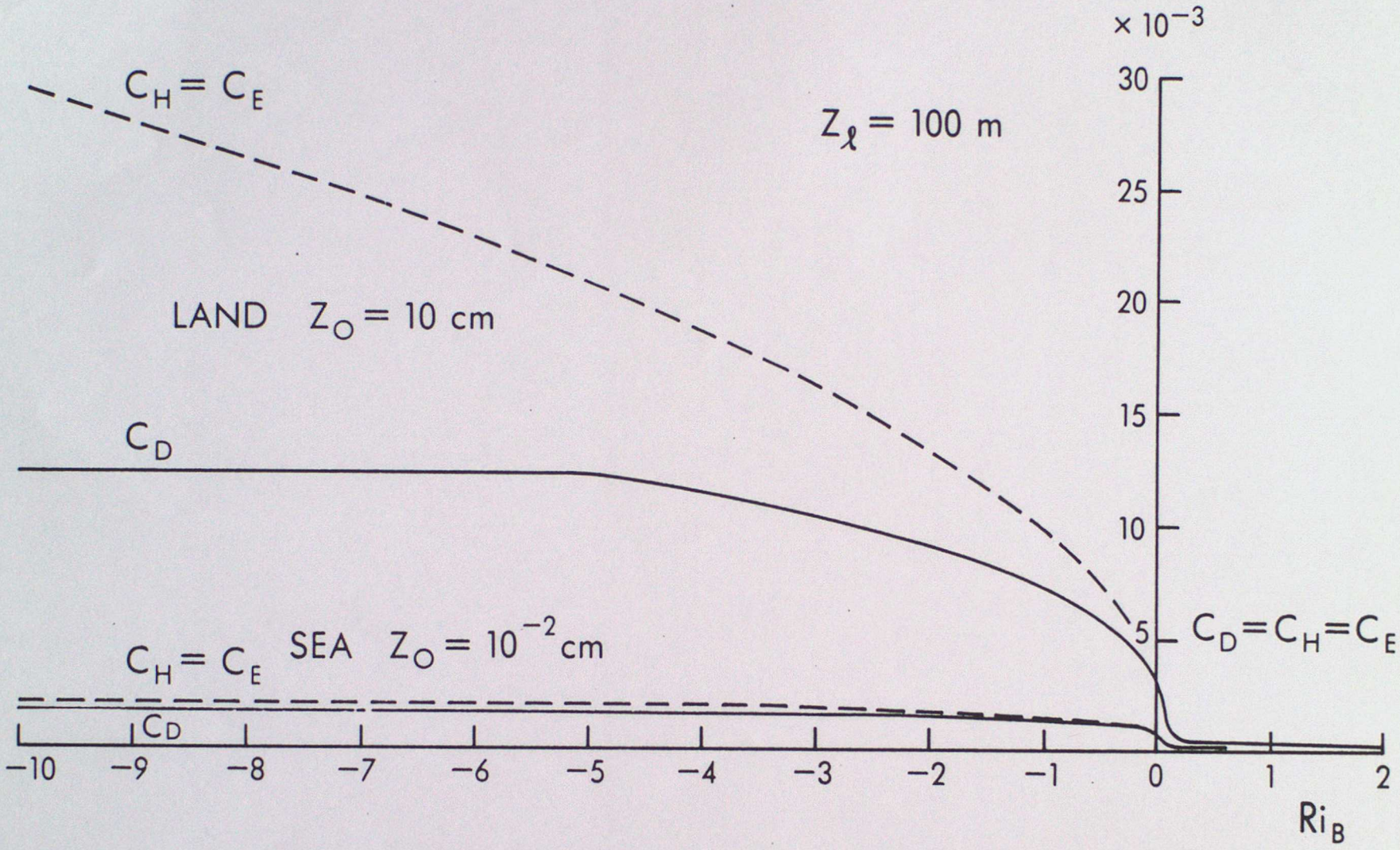


Figure 13 Surface layer bulk transfer coefficients derived from Monin-Obukhov similarity theory and used in the 11-Layer Model. (AGCM)

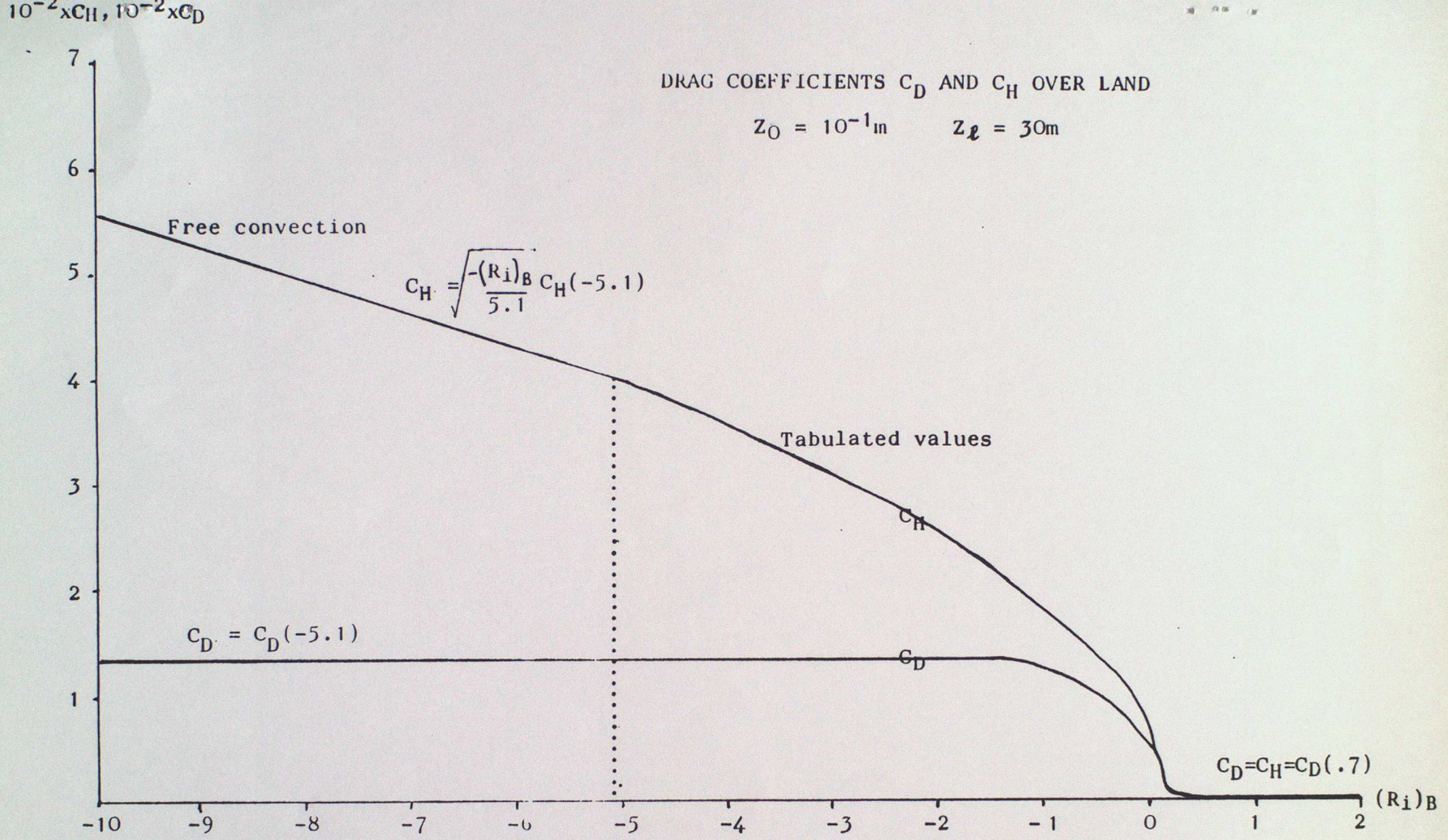


Figure 14. Drag and surface exchange coefficients used over land in the operational model.
$C_E \equiv C_H$ (GWPM and RWPM)

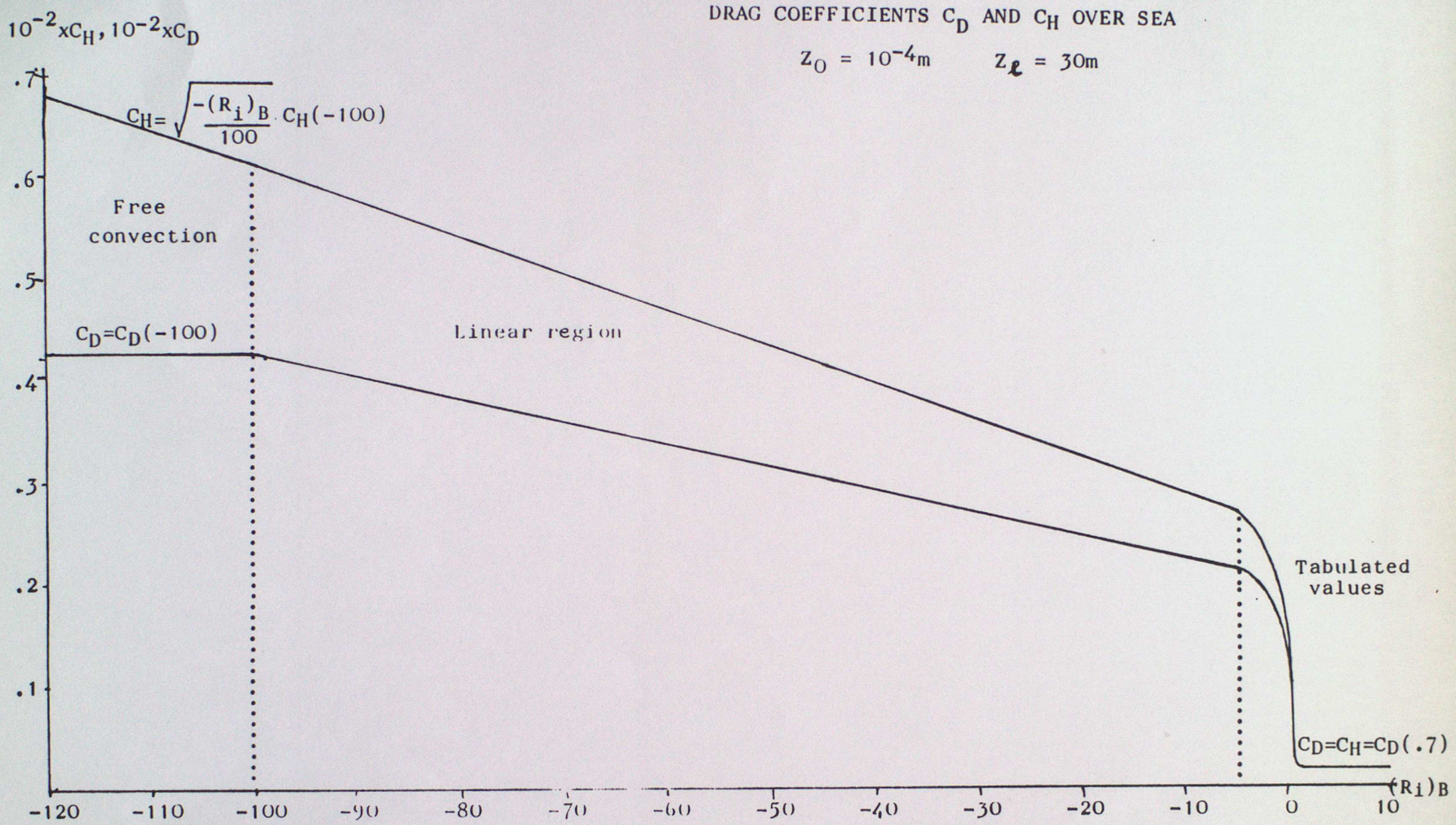
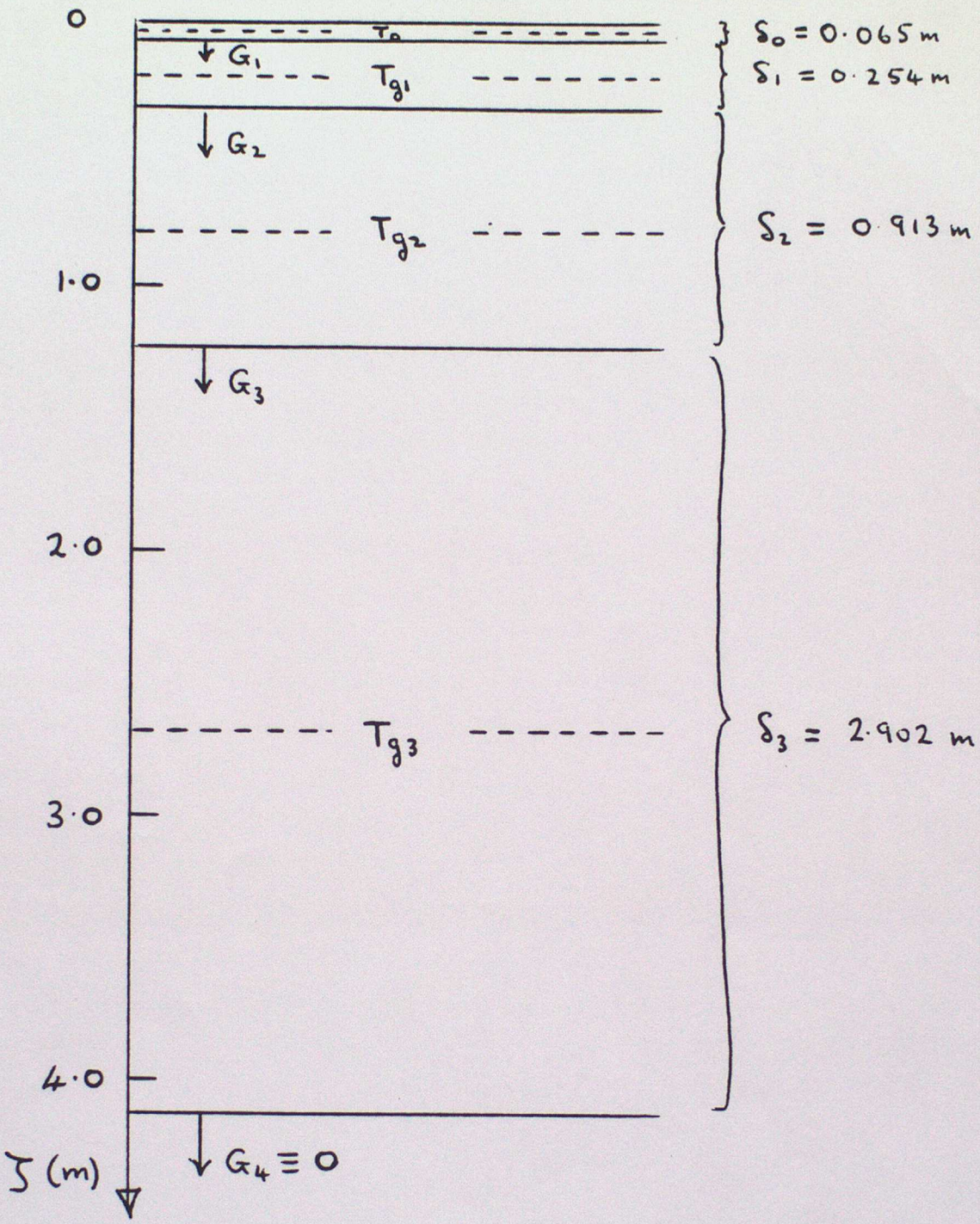


Figure 15. Drag and surface exchange coefficients used over sea in the operational model. (GWPM and RWPM)
$C_E \equiv C_H$

$$G_o = R_N - H - Q - Q_f \quad [\text{Eqn.}(17)]$$



$$\frac{\partial T_g}{\partial t} = -\frac{1}{C}\frac{\partial G}{\partial z}$$

$$G = -\lambda \frac{\partial T_g}{\partial z}$$

$$C = 1.6 \times 10^6 \ J\,m^{-3}K^{-1} \ ; \ \lambda = 1.2 \ W\,m^{-1}K^{-1}$$

Figure 16. Schematic representation of the 4-layer soil-temperature model, due to Warrilow (1986a), used in the AGCM. A version of this model is also used in the RWPM (see Table 2).

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